CHAPTER SEVEN

Joints and Veins

7.1 INTRODUCTION

Visitors from around the world trek to Arches National Park in southeastern Utah (USA) to marvel at its graceful natural arches. These arches appear to have been carved through high, but relatively thin, free-standing sandstone walls. From the air, you can see that the park contains a multitude of such walls, making its landscape resemble a sliced-up loaf of bread (Figure 7.1a). The surfaces of rock walls in Arches Park initiated as **joints,** which are natural fractures in rock across which there has been no shear displacement (see Table 7.1 for a more formal definition). Erosive processes through the ages have preferentially attacked the walls of the joints, so that today you can walk in the space between the walls. Though joints are not always as dramatic as those in Arches National Park, nearly all outcrops contain joints. At first glance, joints may seem to be simple and featureless geologic structures, but in fact they are well worth studying, not only because of their importance in controlling landscape morphology, but also because they profoundly affect rock strength, influence hydrologic properties (such as permeability), and because they can provide a detailed, though subtle history of stress and strain in a region.

Although the basic definition of the term joint is nongenetic, most contemporary geologists believe that they form during Mode I loading (see Chapter 6); that is, that they are tensile fractures that form perpendicular to the σ_3 trajectory and parallel to the principal plane of stress that contains the σ_1 and σ_2 directions. Not all geologists share this viewpoint, and some researchers use the term "joint" when referring to shear fractures as well. This second usage is discouraged, because structures that are technically faults might also be referred to as joints, so we do not use the term "joint" in reference to a shear fracture.

(a)

(b)

FIGURE 7.1 Examples of joints and veins on different scales. (a) Air photo of regional joints in sandstone near Arches National Park, Utah (USA). Note the Colorado River for scale. (b) Veining in limestone exposed in a road cut near Catskill (New York State, USA).

In this chapter, we begin by describing the morphology of individual joints and the geometric characteristics of groups of joints. Then, we discuss how to study joints in the field, and how to interpret them. We conclude by describing **veins,** which are fractures filled with minerals that precipitated from a fluid (Figure 7.1b). But before we begin, we offer a note of caution. The interpretation of joints and veins remains quite controversial, and it is common for field trips that focus on these structures to end in heated debate. As you read this chapter, you'll discover why.

TABLE 7.1 JOINT TERMINOLOGY

Arrest line	An arcuate ridge on a joint surface, located at a distance from the origin, where the joint front stopped or paused during propagation of the joint; also *rib marks.*
Columnar joints	Joints that break rock into generally hexagonal columns; they form during cooling and contraction in hypabyssal intrusions or lava flows.
Conjugate system	Two sets of joints oriented such that the dihedral angle between the sets is approximately 60°.
Continuous joints	Throughgoing joints that can be traced across an outcrop, and perhaps across the countryside.
Cross joints	Discontinuous joints that cut across the rock between two systematic joints, and are oriented at a high angle to the systematic joints.
Cross-strike joints	Joints that cut across the general trend of fold hinges in a region of folded rocks (i.e., the joints cut across regional bedding strike).
Dessication cracks	Joints formed in a layer of mud when it dries and shrinks; dessication cracks (or *mud cracks*) break the layer into roughly hexagonal plates.
Discontinuous joints	Short joints that terminate within an outcrop, generally at the intersection with another joint.
En echelon	An arrangement of parallel planes in a zone of fairly constant width; the planes are inclined to the borders of the zone and terminate at the borders of the zone. In an *en echelon* array, the component planes are of roughly equal length.
Exfoliation	See sheeting joints.
Hackle zone	The main part of a plumose structure, where the fracture surface is relatively rough due to microscopic irregularities in the joint surface formed when the crack surfaces get deflected in the neighborhood of grain-scale inclusions in the rock, or due to off-plane cracking (formation of small cracks adjacent to the main joint surface) as the fracture propagates.
Hooking	The curving of one joint near its intersection with an earlier formed joint.
Inclusion	A general term for any solid inhomogeneity (e.g., fossil, pebble, burrow, xenolith, amygdule, coarse grain, etc.) in a rock; inclusion may cause local stress concentrations.
Joint	A natural, unfilled, planar or curviplanar fracture which forms by tensile loading (i.e., the walls of a joint move apart very slightly as the joint develops). Joint formation does not involve shear displacement.

7.2 SURFACE MORPHOLOGY OF JOINTS

If you look at an exposed joint surface, you'll discover that the surface is not perfectly smooth. Rather, joint surfaces display a subtle roughness that often resembles the imprint of a feather. This pattern is called a plumose structure (Figure 7.2).

7.2.1 Plumose Structure

Plumose structures form at a range of scales, depending on the grain size of the host rock. In very fine-grained coal, for example, components of plumose structure tend to be much smaller than in relatively coarser siltstone. Some of the best examples of plumose structure form in fine-grained rocks like shale, siltstone, and basalt, but you might not see obvious plumose structure on joints in very coarse-grained rocks like granite. Let's look at plumose structure a little more closely (Figure 7.3a). A plumose structure spreads outward from the **joint origin,** which, as the name suggests, represents the point at which the joint started to grow. Joint origins typically look like small dimples in the fracture plane (Figure 7.3b). Several distinct morphologies surround the joint origin. In the **mirror zone,** which lies closest to the origin, the joint surface is very smooth. Further from the origin, the mirror zone merges with the **mist zone,** in which the joint surface slightly roughens. Mirror and mist zones, while they are well developed in joints formed in

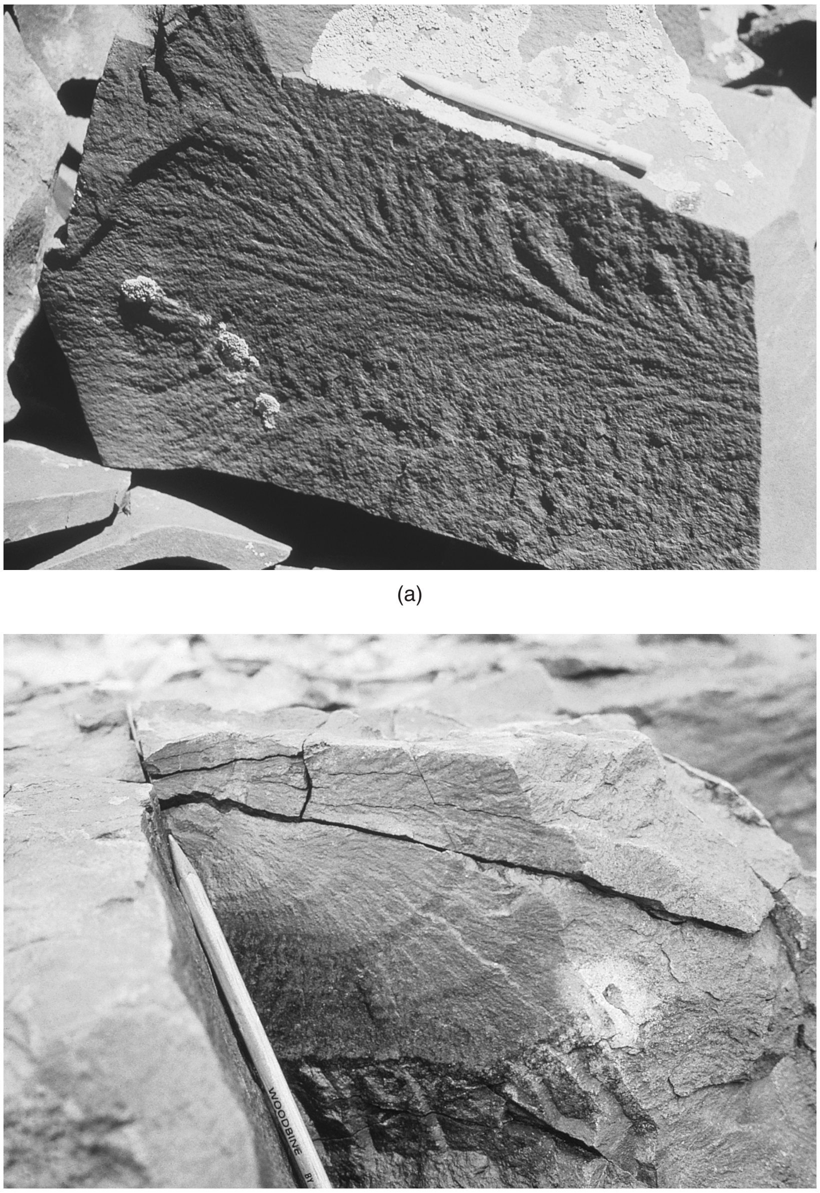
TABLE 7.1 JOINT TERMINOLOGY (CONTINUED)

Joint array	Any group of joints (systematic or nonsystematic).
Joint density	The surface area of joints per unit volume of rock (also referred to as *joint intensity*).
Joint origin	The point on the joint (usually a flaw or inclusion) at which the fracture began to propagate; it is commonly marked by a dimple.
Joint set	A group of systematic joints.
Joint stress shadow	The region around a joint surface where joint-normal tensile stress is insufficient to cause new joints to form.
Joint system	Two or more geometrically related sets of joints in a region.
Mirror region	Portion of a joint surface adjacent to the joint origin where the surface is very smooth; mirrors do not occur if the rock contains many small-scale heterogeneities.
Mist region	A portion of a joint surface surrounding the mirror where the fracture surface begins to roughen.
Nonsystematic joints	A joint that is not necessarily planar, and is not parallel to nearby joints.
Orthogonal system	Two sets of joints that are at right angles to one another.
Plume axis	The axis of the plume in a plumose structure.
Plumose structure	A subtle roughness on the surface of some joints (particularly those in fine-grained rocks) that macroscopically resembles the imprint of a feather.
Sheeting joints	Joints formed near the ground surface that are roughly parallel to the ground surface; sheeting joints on domelike mountains make the mountains resemble delaminating onions; also *exfoliation.*
Strike-parallel joints	Joints that parallel the general trend of fold-hinges in a region of folded strata (i.e., the joints parallel regional bedding strike).
Systematic joints	Roughly planar joints which occur as part of a set in which the joints parallel one another, and which are relatively evenly spaced from one another.
Twist hackle	One of a set of small *en echelon* joints formed along the edge of a larger joint; a twist hackle is not parallel to the larger joint, and forms when the fracture surface twists continuously into a different orientation and then breaks up into segments.

glassy rocks, are difficult to recognize in coarser rocks. Continuing outward, the mist zone merges with the **hackle zone,** in which the joint surface is even rougher. It is the hackle zone that forms most of the plumose structure. Roughness in the hackle zone defines vague lineations, or **barbs,** that curve away from a **plume axis,** which together comprise the feather-like plume. The acute angle between the barbs and the axis points back toward the joint origin, so the plume defines the local direction of joint propagation. The median line may be fairly straight and distinct, or it may be wavy and diffuse (Figure 7.4a and b). On some joint surfaces, concentric ridges known as **arrest lines** (Figure 7.4c) form on the joint surface at a distance from the origin. These ridges represent, as the name indicates, breaks in the growth of the joint.

7.2.2 Why Does Plumose Structure Form?

Mode I loading of a perfectly isotropic and homogeneous material should yield a perfectly smooth, planar fracture that is oriented perpendicular to the remote σ_3. Real joints are not perfectly smooth for two reasons. First, real rocks are not perfectly isotropic and homogeneous, meaning that the material properties of a rock change from point to point in the rock. Inhomogeneities exist because not all grains in a rock have the same composition and because not all grains are in perfect contact with one another. The presence of inhomogeneities distorts the local stress field at the tip of a growing joint, so that the principal stresses at the tip are not necessarily parallel to the remote σ_3. As a



(a) (b)

FIGURE 7.2 Photographs of plumose structure on joint surfaces (New York State, USA). (a) Wavy plumose structure on a joint in siltstone. (b) Plumose structure in thin bedded siltstone; pencil points to the point of origin.

consequence, the joint-propagation path slightly twists and tilts as the joint grows.

Second, the stress field at the tip of a crack changes as the crack tip propagates. Recall from Equation 6.4 that the stress intensity at the crack tip is proportional to the length of the crack, and that the magnitude of the local tensile stress at the tip of the crack is, in turn, proportional to the stress intensity. Thus, as the crack grows, the stress intensity at the crack tip grows, up to a limiting value. Experimental work demonstrates that the velocity of crack-tip propagation is also proportional to the stress intensity. Stress magnitude and tip-propagation velocity are relatively small near the joint origin, because the crack is very short, and increase with distance from the origin, eventually reaching a maximum (called terminal velocity). If the stress magnitude at the tip exceeds a critical value, the energy available for cracking rock exceeds the energy needed to create a single surface. The excess energy goes into breaking bonds off the plane of the main joint surface, resulting in the formation of microscopic cracks that splay off the main joint. If the energy becomes very large, the crack may actually split into two separate, parallel surfaces.

With these two conditions in mind, we can now explain why plumose structure has distinct morphological features. The dimple at the origin forms because the flaw[1] at which the joint nucleated either was not perpendicular to the remote σ_3, or caused a local change in the orientation of stress trajectories. The portion of the joint that formed in the immediate vicinity of the origin was, therefore, not perpendicular to the remote σ_3. As soon as the crack propagated away from the flaw, it curved into parallelism with the $\sigma_1\sigma_2$ principal plane (Figure 7.3). In the mirror zone, the joint is still short, so the stress intensity, tensile stress magnitude, and tip-propagation velocity are all relatively small. As a consequence of the low stress, only bonds in the plane exactly perpendicular to the local σ_3 can break, so the joint surface that forms is very smooth. In the mist zone, however, the joint moves faster, stress is higher, and stress at the joint tip is sufficiently large to break off-plane bonds, thereby forming microscopic off-plane cracks that make the surface rougher than in the mirror zone. In the hackle zone, the joint tip is moving at its terminal velocity and stresses at the crack tip are so large that larger off-plane cracks propagate and the crack locally bifurcates at its tip to form microscopic splays that penetrate the joint walls. The roughness of the hackle zone also reflects the formation of tiny splays

[1]Flaws at which joints initiate include open pores, preexisting microcracks, irregularities on a bedding plane, inclusions (like a pebble, fossil, amygdule, or concretion), or primary sedimentary structures (a sole mark or ripple).

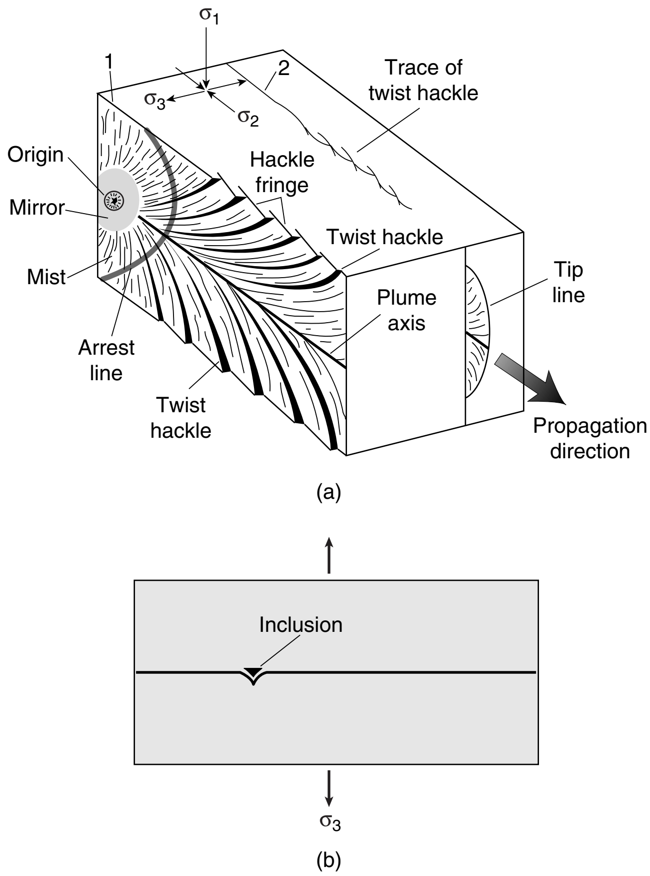


FIGURE 7.3 (a) Block diagram showing the various components of an ideal plumose structure on a joint. The face of joint 1 is exposed; joint 2 is within the rock. (b) Simple cross-sectional sketch showing the dimple of a joint origin, controlled by an inclusion.

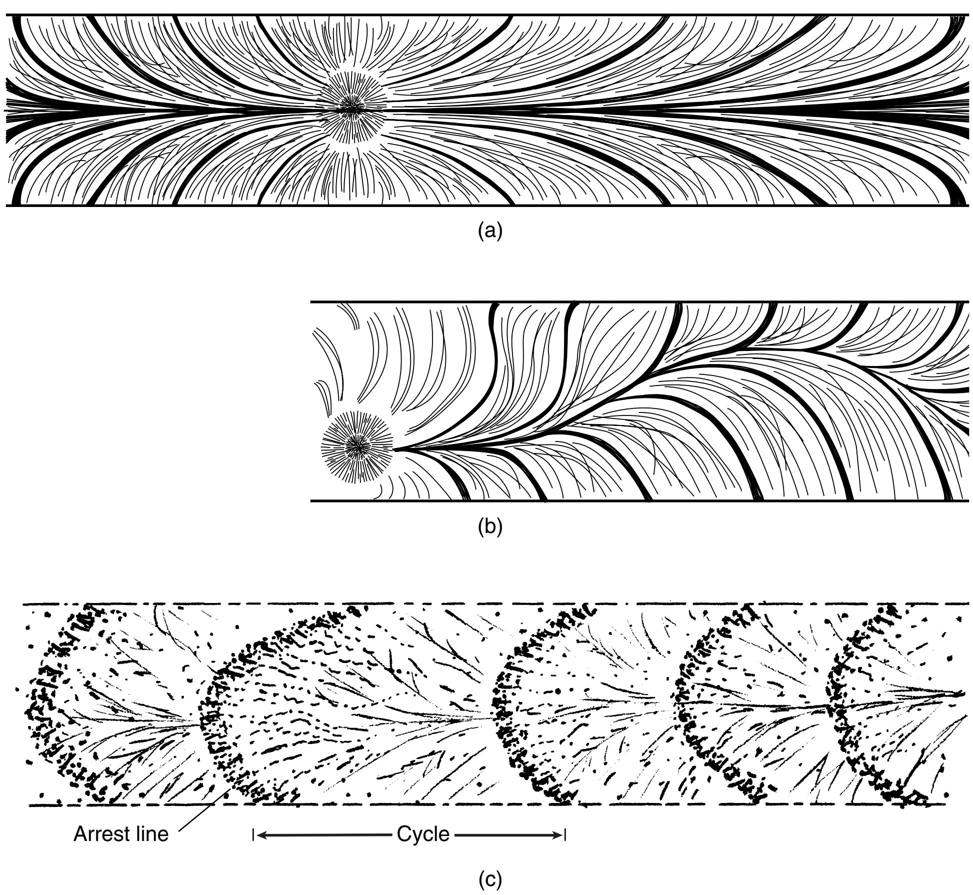


FIGURE 7.4 Types of plumose structure. (a) Straight plume. (b) Curvy plume. (c) Plume with many arrest lines, suggesting that it opened repeatedly.

and warps of the joint surface where the joint tip twists or tilts as it passes an inclusion and breaks into microscopic steps. Arrest lines on a joint surface represent places where the fracture tip pauses between successive increments of propagation. The visible ridge of the arrest line, in part, represents the contrast between the rough surface of the hackle and the relatively smooth surface of the mirror/mist zone formed as the fracture begins to propagate, and in part may be analogous to the dimple formed at a crack origin. Thus, plumose structures form because of the twisting, tilting, and splitting occurring at the tip due to variations in local stress magnitude and orientation.

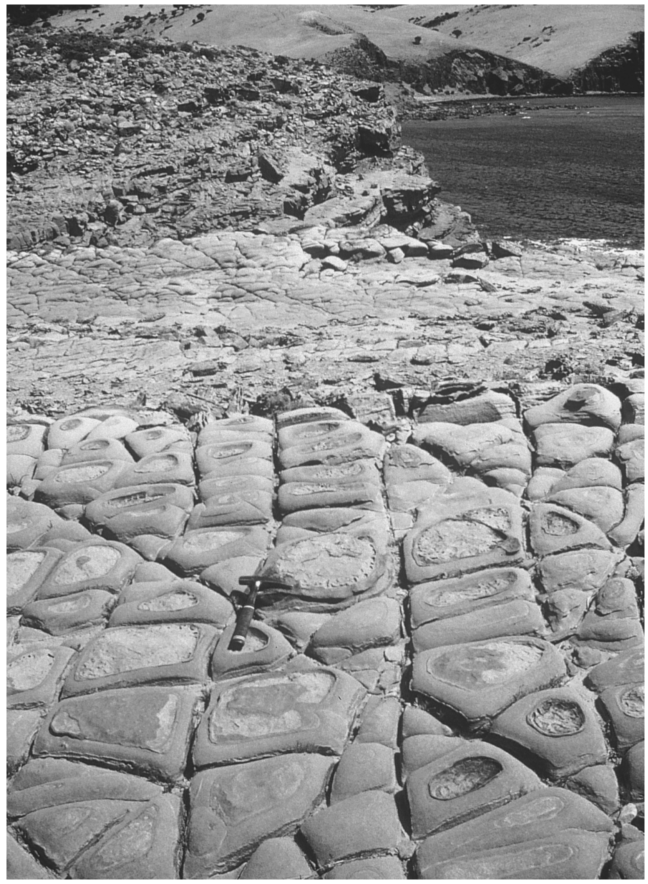
7.2.3 Twist Hackle

Features such as bedding planes and preexisting fractures locally modify the orientation of principal stresses because they approximate free surfaces.[2] If a growing joint enters a region where it no longer parallels a principal plane of stress (for example, as occurs when the crack tip of a joint in a sedimentary bed approaches the bedding plane), the crack tip pivots to a new orientation. As a consequence, the joint splits into a series of small *en echelon* joints, because a joint surface cannot twist and still remain a single continuous surface. The resulting array of fractures is called **twist hackle,** and the edge of the fracture plane where twist hackle occurs is called the **hackle fringe** (Figure 7.3a). Note that if the hackle fringe intersects an outcrop face, the trace of a large planar joint within the outcrop may look like a series of small joints in an *en echelon* arrangement.

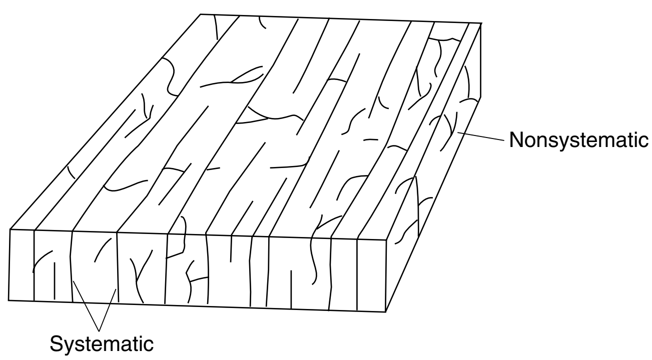
7.3 JOINT ARRAYS

7.3.1 Systematic versus Nonsystematic Joints

Systematic joints are planar joints that comprise a family in which all the joints are parallel or subparallel to one another, and maintain roughly the same average spacing over the region of observation (Figure 7.5). Systematic joints may cut through many layers of strata, or be confined to a single layer. **Nonsystematic joints** have an irregular spatial distribution, they do not parallel neighboring joints, and they tend to be nonplanar (Figure 7.5b). Nonsystematic joints may terminate at other joints. You will often find both systematic and nonsystematic joints in the same outcrop.



(a)



(b)

FIGURE 7.5 (a) Three sets of systematic joints controlling erosion in Cambrian sandstone (Kangaroo Island, Australia). (b) Block diagram showing occurrence of both systematic and nonsystematic joints in a body of rock.

[2]A "free surface" is a surface across which there is no cohesion, so it cannot transmit shear stresses. By definition, a free surface is a principal plane of stress, but if the free surface is not parallel to a principal plane of the remote stress, then the remote stress trajectories change orientation so that they are either parallel or perpendicular to the free surface.

7.3.2 Joint Sets and Joint Systems

Describing groups of joints efficiently requires a fair bit of jargon. Matters are made even worse because not all authors use joint terminology in the same way, so it's good practice to define your terminology in context. We'll describe joint patterns here and give the explanations of why various different groups of joints form later in the chapter.

A **joint set** is a group of systematic joints. Two or more joint sets that intersect at fairly constant angles comprise a **joint system,** and the angle between two joint sets in a joint system is the **dihedral angle.** If the two sets in a system are mutually perpendicular (i.e., the dihedral angle is ~90°), we call the pair an **orthogonal system** (Figure 7.6a), and if the two sets intersect with a dihedral angle significantly less than 90° (e.g., a dihedral angle of 30° to 60°), we call the pair a **conjugate system** (Figure 7.6a). Many geologists use the terms "orthogonal" or "conjugate" to imply that the pair of joint sets formed at the same time. However, as you will see later in this chapter, nonparallel joint sets typically form at different times. So, we use the terms merely to denote a geometry, not a mode or timing of origin.

As shown in Figure 7.6a, many different configurations of joint systems occur, which are distinguished from one another by the nature of the intersections between sets and by the relative lengths of the joints in the different sets. In joint systems where one set consists of relatively long joints that cut across the outcrop whereas the other set consists of relatively short joints that terminate at the long joints, the throughgoing joints are **master joints,** and the short joints that occur between the continuous joints are **cross joints** (Table 7.1).

In the flat-lying sedimentary rocks that occur in continental interior basins and platforms (e.g., the Midwest region of the United States), joint sets are

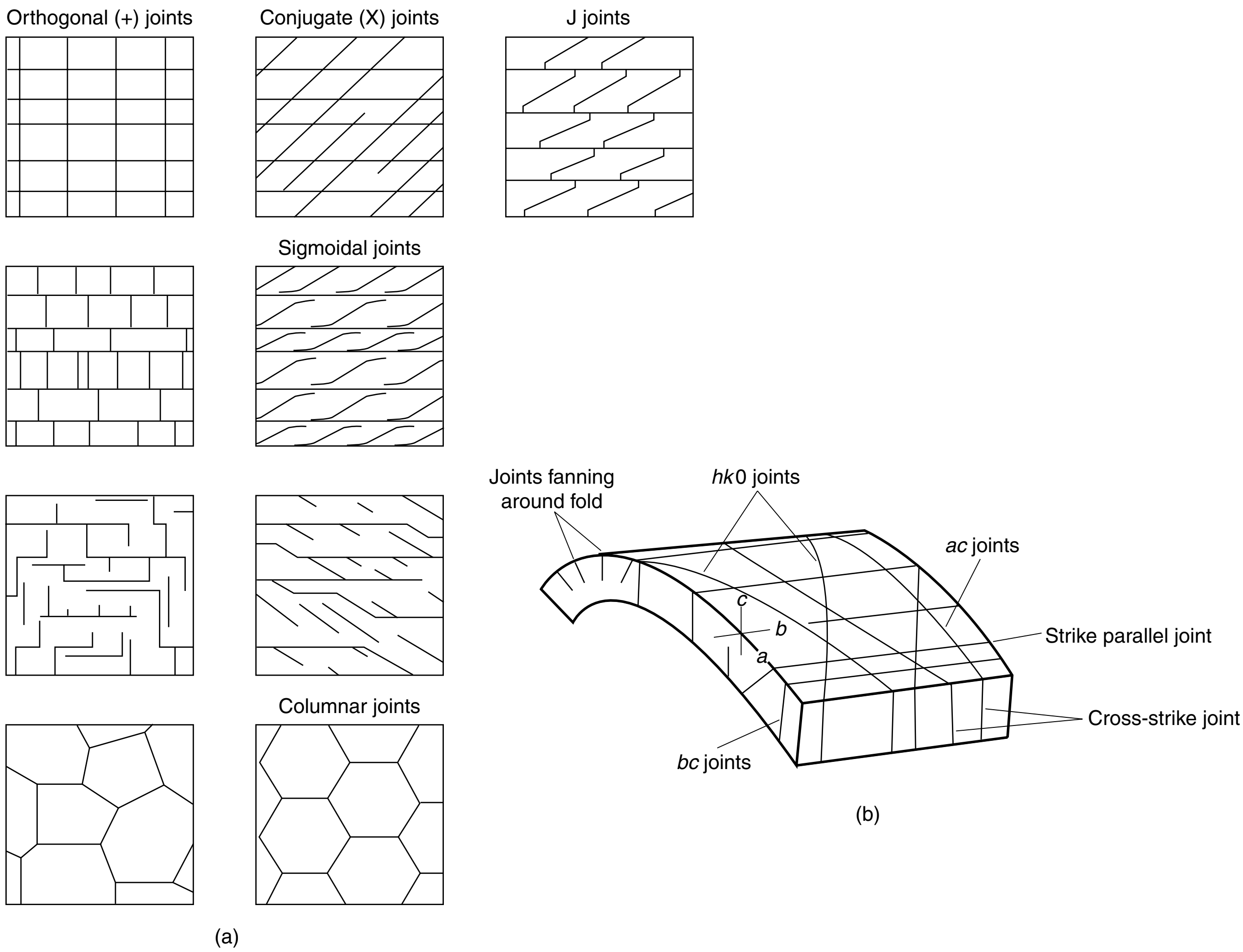


FIGURE 7.6 (a) Traces of various types of joint arrays on a bedding surface. (b) Idealized arrangement of joint arrays with respect to fold symmetry axes. The "*hk*0" label for joints that cut diagonally across the fold-hinge is based on the Miller indices from mineralogy; they refer to the intersections of the joints with the symmetry axes of the fold.

perpendicular to the ground surface (and, therefore, to bedding) and orthogonal systems are common. In gently folded sedimentary rocks, such as along the foreland margin of a mountain range (e.g., the western side of the Appalachians), strata contain both vertical joint sets that cut across the folded layers, and joints that are at a high angle to bedding and fan around the folds (Figure 7.6b). Both orthogonal and conjugate systems occur in such gently folded strata. The joint sets of an orthogonal system in folded sedimentary rocks commonly have a spatial relationship to folds of the region, so we can distinguish between **strike-parallel joints,** which parallel the general strike of bedding (roughly parallel to regional fold hinges), and **cross-strike joints,** which trend at high angles (~60° to 90°) to the regional bedding strike (Figure 7.6b).[3] Conjugate systems in gently folded rocks consist of two cross-strike sets with their acute bisector at a high angle to the fold hinge. Because both sets of joint systems need not form at the same time, a conjugate geometry of a system of joints does not require that they are conjugate shear ruptures.

In the internal portions of mountain belts, where rocks have been intensely deformed and metamorphosed, outcrops may contain so many joints that joint systems may be difficult to recognize or simply do not exist. In such regions, joints formed prior to deformation and metamorphism have been partly erased. New joints then form at different times during deformation, during subsequent uplift, or even in response to recent stress fields. Rocks in such regions are so heterogeneous that the stress field varies locally, and thus joints occur in a wide range of orientations. Nevertheless, in some cases, younger joints, meaning those formed during uplift or due to recent stress fields, may stand out as distinct sets.

Intrusive and metamorphic rocks without a strong schistosity (such as granite, migmatitic gneiss) commonly contain a set of joints that roughly parallels ground surface topography, and whose spacing decreases progressively toward the surface. Such joints are called **sheeting joints** or **exfoliation joints** (Figure 7.7). If the ground surface is not horizontal, as is the case on the sloping side of a mountain, sheeting joints curve and follow the face of the mountain, thus giving the mountain the appearance of a partially peeled onion. Rock sheets detach off the mountain along these joints, thereby creating smooth dome-shaped structures known as **exfoliation domes.** Half Dome, a challenge that draws mountain climbers to Yosemite National Park in the Sierra Nevada Mountains of California, is an exfoliation dome, one half of which was cut away by glacial erosion.

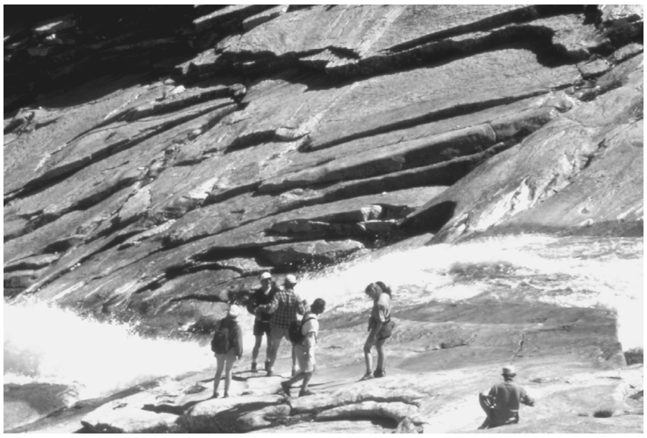


FIGURE 7.7 Sheeting joints (or exfoliation) in granite of the Sierra Nevada.

Shallow intrusive igneous rock bodies (dikes and sills) and lava flows in many localities display **columnar jointing,** meaning that they have been broken into joint-bounded columns which, when viewed end-on, have roughly hexagonal cross sections (Figure 7.6a). In the case of sheet intrusions, the long axes of the columns tend to be perpendicular to the boundaries of the sheet (horizontal in dikes and vertical in sills); however, in some bodies the columns curve. The visual impression of columnar jointing catches people's imaginations, so these structures tend to be dubbed with unusual names like Giant's Causeway (in Ireland), Devil's Postpile (in California; see Figure 2.25), and the Spielbergian platform of Devil's Tower (in Wyoming).

7.3.3 Cross-Cutting Relations Between Joints

The way in which nonparallel joints intersect one another provides information concerning their relative ages. For example, if joint A terminates at its intersection with joint B, then joint A is younger, because a propagating fracture cannot cross a free surface, and an open preexisting joint behaves like a free surface.[4]

A younger joint's orientation also may change where it approaches an older joint that behaves like a free surface. Why? Remember that at, or near, a free

[3]Note that "cross-strike joints" are not necessarily the same as "cross joints" (Table 7.1).

[4]Joints that are filled, or whose faces are tightly held together by stress, can transmit at least some shear stress, and thus do not behave like perfect free surfaces.

surface, a Mode I fracture must be either parallel, or perpendicular, to the surface so as to maintain perpendicularity to σ_3. Thus, near a free surface, the local stress field differs from the remote stress field if the free surface does not parallel a principal plane of the remote stress field. If an older joint (joint B) acts as a free surface, then the younger joint (A) curves in the vicinity of joint B to become parallel to the local principal plane of stress adjacent to B, unless it already happens to parallel a principal plane of stress. The way in which the younger joint curves depends on the stress field. If the local σ_3 adjacent to the older joint is parallel to the walls of the older joint, then the younger joint curves so that it is orthogonal to the first joint at their point of intersection, a relationship called **hooking;** such a structure is called a J junction (Figure 7.6a). However, if the local σ_3 is perpendicular to the walls of the older joint, then the younger joint curves into parallelism with joints of the first set, and has a sigmoidal appearance (Figure 7.6a).

In some joint systems, two nonparallel joints appear to cross one another without any apparent interaction; in other words, they are mutually cross cutting. Such intersections are sometimes referred to as "+" intersections, if the joints are orthogonal, or "×" intersections if they are not orthogonal (Figure 7.6a). These relationships may represent situations where (1) the earlier joint did not act as a free surface, (2) the intersection of two younger joints at the same point on an older joint is simply coincidental, or (3) the cross-cutting relationship is an illusion—within the body of the outcrop, the older joint terminated, and the younger joint simply grew around it.

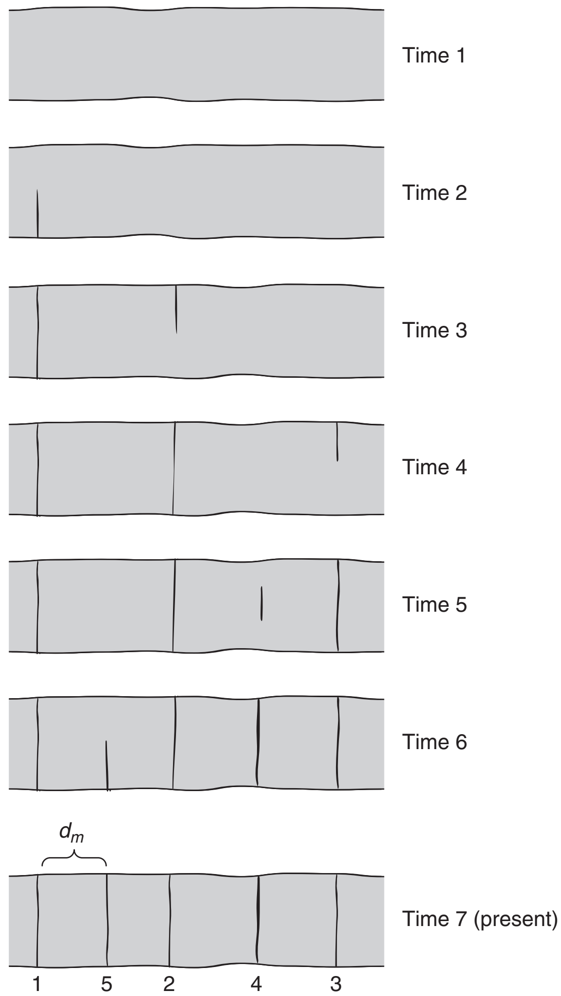


FIGURE 7.8 A model of the sequence of development of joints. Time 1 refers to the time before the first joint forms, and time 7 is the present day. This scenario suggests that joints form in a random sequence, but with regular spacing.

7.3.4 Joint Spacing in Sedimentary Rocks

When looking at jointing in a sequence of stratified sedimentary rock, you might notice that within a bed, joints are often evenly spaced. Where this occurs, we can define **joint spacing** as the average distance between adjacent members of a joint set, measured perpendicular to the surface of the joint. Informally, geologists refer to joints as being "closely spaced" or "widely spaced" in a relative sense, but to be precise, you should describe joint spacing in units of length (e.g., 5 cm).

In order to understand why joints are evenly spaced, we look at how an array of joints develops in a bed. Consider a bed of sandstone that contains five joints (Figure 7.8). Experimental work suggests that joints form in sequence; that is, first joint 1, then joint 2, then joint 3, and so on. When a new joint forms, it is at some distance greater than a minimum distance (d_m) from a preexisting joint. Formation of a joint relieves tensile stress for a critical distance, d_m (Figure 7.9). The zone on either side of a joint in which there has been a decrease in tensile stress is called the **joint stress shadow.** Stresses sufficient to create the next joint are only achieved outside of this shadow, and are created by traction between the bed and beds above and below it, as well as by stress transmitted within the bed beyond the fracture front of the preexisting joint. The spacing between joints is determined by the width of the joint stress shadow; so, because the shadow is about the same width for all joints in the bed, the spacing ends up being fairly constant. Joint spacing depends on four parameters: bed thickness, stiffness, tensile strength, and strain. We'll examine each of these parameters in turn.

Relation between joint spacing and bed thickness. All other parameters being equal, joints are more

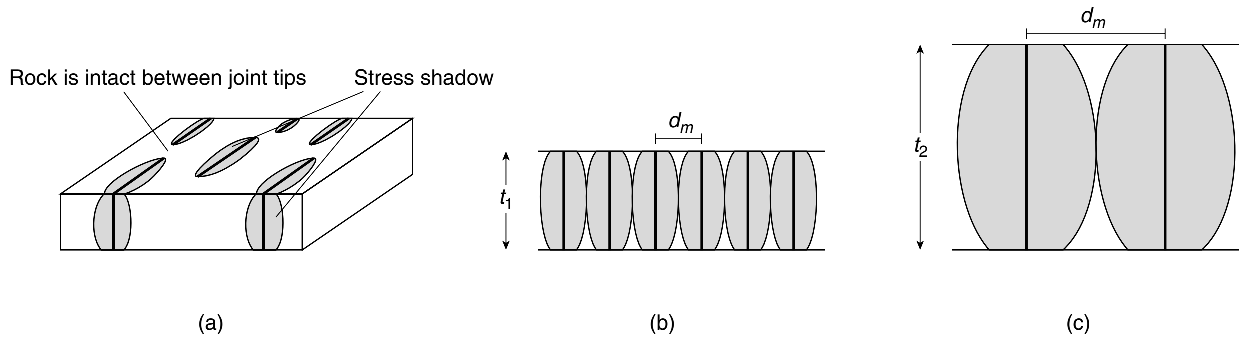


FIGURE 7.9 The concept of stress shadows around joints. The heavy vertical lines are joints; d_m refers to the average spacing between joints. (a) Block diagram illustrating stress shadow (shaded area) around each joint. Note how stress is transmitted across regions that are unfractured in the third dimension. Stresses are also exerted by tractions at bedding contacts. (b) Thin bedded sequence, containing joints with narrow stress shadows, so that the joints are closely spaced. (c) Thick bedded sequence, containing joints with wide stress shadows, so that the joints are widely spaced.

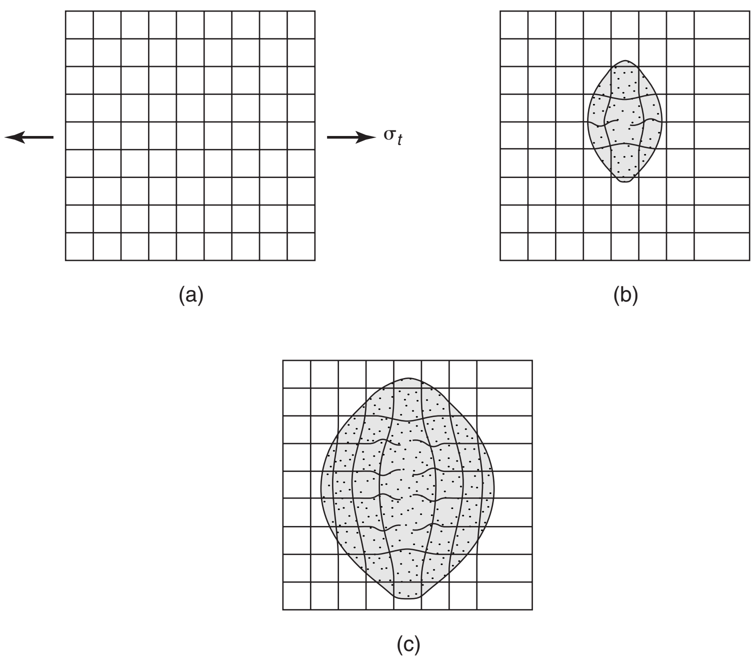


FIGURE 7.10 Illustration of why joint stress shadows exist. (a) A grid of springs. (b) Cutting one spring causes only a few springs to relax around the cut, so only a relatively small area is affected, as indicated. (c) Cutting many springs in a row causes a wider band of springs to relax; thus, a larger area is affected.

closely spaced in thinner beds, and are more widely spaced in thicker beds. The relationship is a reflection of joint-stress shadow width, because the greater the length of the joint (i.e., length of the joint trace in a plane perpendicular to bedding and joint), the wider the stress shadow (Figure 7.9b and c). To picture why this is so, imagine a net composed of springs (Figure 7.10a). If you reach into the net and cut one spring, only a few of the neighboring springs relax (Figure 7.10b); however, if you cut many of the springs in a row, a much wider zone of neighboring springs relaxes (Figure 7.10c). In thicker beds, joint stress shadows are wider, so joints tend to be more widely spaced.

Relation between joint spacing and lithology. Recall that the *stiffness* (i.e., the elastic value *E*, Young's Modulus) of a rock layer depends on lithology (i.e., Hooke's law states that $\sigma = E \cdot \mathbf{e}$; see Chapter 5). Imagine a block of rock composed of sandstone and dolomite (Figure 7.11). Dolomite is stiffer ($E \approx -600$ MPa) than sandstone ($E \approx -200$ MPa). We stretch the block under brittle conditions by a uniform amount so that all layers undergo exactly the same elongation (**e**). The stress that develops in each bed is defined by Hooke's law; however, since the elongation is the same for each bed, the magnitude of σ depends on *E*. Thus, beds composed of rock with a larger *E* develop a greater stress and fracture first. In the model of Figure 7.11, the stiffer dolomite bed probably fractured a few times before the sandstone bed fractured for the first time, so more joints develop in the dolomite bed than in the sandstone bed. In sum, for a given strain, stress is larger in stiffer beds, so other factors being equal, stiffer beds have smaller joint spacings.

Relation between joint spacing and tensile strength. Predicting fracture spacing cannot be done by considering *E* alone, because, in some cir-

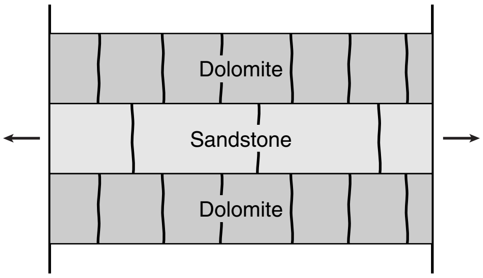


FIGURE 7.11 Cross-sectional sketch illustrating a multilayer that is composed of rocks with different values of Young's modulus. The stiffer layers (dolomite) develop more closely spaced joints.

cumstances, a rock with a smaller E may actually have a lower tensile strength than a rock with a larger E. Thus, it will crack at a lower strain than a rock with a larger E, if the rock with the larger E also has a larger tensile strength. Other factors being equal, rocks with smaller tensile strength develop more closely spaced joints.

Relation between joint spacing and the magnitude of strain. A bed that has been stretched more contains more joints than a bed that has been stretched less, as you might expect.

Overall, the spacing of well-developed joints is about equal to the bed thickness. If you ever have the chance to hike down the Grand Canyon, don't forget to look at the jointing in different units as you descend. Bedding planes tend to be weak and do not transmit shear stress efficiently, so joints typically terminate at bedding planes. Because joint spacing depends on bed thickness and lithology, joint spacing varies from bed to bed. Weak, thinly bedded shales contain such closely spaced joints that they break into tiny fragments, and, as a consequence, they tend to form slopes. In contrast, thick sandstone beds develop only widely spaced fractures, so thick sandstone beds typically protrude and hold up high cliffs.

7.4 JOINT STUDIES IN THE FIELD

Before discussing how to go about studying joints in the field, it's worth discussing *why* you might want to study joints in the field. Perhaps the most common reasons to study joints are for engineering or hydrologic applications. As we noted before, fractures affect the strength of foundations, quarrying operations, excavations, groundwater and (toxic) waste flow, and slope stability. For example, if you find that a region contains a systematic joint set that is oriented north–south, you can expect groundwater to flow faster in the north–south direction than in the east–west direction, or that quarrying might be easier if the quarry walls strike north–south than if they strike east–west. But the study of jointing has applications to more academic geologic issues as well. Geologists who are interested in tectonics study joints to see if they provide information about (paleo)stress fields, and geomorphologists study joints to find out if they control the drainage patterns or the orientation of escarpments. With these goals in mind, what specifically do we look for in a joint study? In most cases the questions that we ask include the following:

1. Is jointing in the outcrop systematic or nonsystematic? In other words, can we define distinct sets of planar joints and/or regularly oriented cross joints in an outcrop, or does the outcrop contain irregular and randomly oriented joints with relatively short traces (i.e., nonsystematic joints)? If nonsystematic jointing is present, is it localized or pervasive? Formulating hypotheses on joint formation in a region needs to take into account whether the joints are systematic or not. Systematic joints likely reflect regional tectonic stress trajectories at the time of fracturing, whereas nonsystematic joints reflect local heterogeneities of the stress field. While nonsystematic joints may be important for determining rock strength and permeability, they provide no information on regional paleostress orientation.
2. What are the orientations of joint sets, if present? If there is more than one set, is there a consistent angular relationship between them, such that we can describe a joint system? Information on the orientation and distribution of joint sets and systems is critical for engineering and hydrologic analyses. For example, joint sets that run parallel to a proposed road cut would create a greater rockfall hazard than joints that are perpendicular to the cut.
3. What is the nature of cross-cutting relationships between joints of different sets, and what is the geometry of joint intersections? Do joints cross without appearing to interact, do they curve to create J intersections, or do they curve into parallelism with one another? Knowledge of cross-cutting relations allows us to determine whether one set of joints is older or younger than another set of joints, a determination that is critical to tectonic interpretations using joints.

4. What is the surface morphology of the joints? Is plumose structure visible on joint surfaces, and if so, what types of plumes (wavy or straight) are visible? Are numerous arrest lines clearly evident on the joint surface? The presence of plumose structure is taken as proof that a joint propagated as a Mode I fracture, and the geometry of the plume provides clues as to the way in which the joint propagated (e.g., in a single pulse, or in several distinct pulses). If a joint surface contains numerous origins, it probably initiated at different times along its length. Joints whose surfaces contain many arrest lines probably propagated in increments. Later we will see that starts and stops of a fracture may indicate that growth was controlled by fluctuating fluid pressure in the rock. Are there other structural features superimposed on joints (e.g., stylolitic pits or slip lineations)? If joints display surface features other than plumose structure, the features indicate post–joint formation strain. Stylolitic pitting on a joint indicates compression and resulting pressure solution (see Chapter 9) across the joint, and slip lineations suggest that the joint was reactivated as a fault later in its history.
5. What are the dimensions of joints? In other words, are the trace lengths of joints measured in centimeters or hundreds of meters? The effect that jointing has on rock strength and rock permeability over a region is significantly affected by the dimensions of the joints. For example, the presence of large throughgoing joints that parallel an escarpment contribute to the hazard of escarpment collapse more than will short nonsystematic joints. Tectonic geologists commonly focus attention on interpretation of large joints, on the assumption that these reflect regional tectonic stress conditions, which are of broad interest.
6. What is the spacing and joint density in outcrop? By joint spacing, we mean the average distance between regularly spaced joints. Information on joint spacing provides insight into the mechanical properties of rock layers and their fracture permeability. By joint density, we mean, in two dimensions, what is the trace length of joints per unit area of outcrop, or in three dimensions, what is the area of joints per unit volume of outcrop? Joint density depends both on the length of the joints and on their spacing. Information on joint density helps define the fracture-related porosity and permeability of a rock body.
7. How is the distribution of joints affected by lithology? In sedimentary rocks, do individual joints cut across a single bed, or do they cut across many beds, or even through the entire outcrop and beyond? In what way is joint spacing affected by bed composition? In the igneous rocks and their contact zones, is the joint spacing or style (both within an igneous body and in the country rock that was intruded) controlled by the proximity of the joint to contact? Information about the relationship between jointing and lithology can be related to physical characteristics such as Young's modulus (E), and can help determine variations in fracture permeability as a function of position in a stratigraphic sequence. Information on the relationship between jointing and lithology may also give insight into the cause of joint formation.
8. Are joints connected to one another or are they isolated? The connectivity of joints is critical to a determination of whether they could provide a permeable network through which fluids (for example, contaminated groundwater or petroleum) could flow.
9. How are joints related to other structures and fabrics? Are joints parallel to tectonic foliations? Are joints geometrically related to folds? Are joints reoriented by folding, or do they cut across the folds? Is there a relationship between joint orientation and measured contemporary stresses? Is the spacing or style of joints related to the proximity of faults? Information on the relation of joints to other structures provides insight into the tectonic conditions in which joints form and the timing of joint formation with respect to the formation of other structures in a region.

7.4.1 Dealing with Field Data about Joints

There are basically two ways to carry out a field study of joint orientation, spacing, and intensity. In the **inventory method,** you define a representative region and measure all joints that occur within that region. For example, you draw a circle or square on the outcrop and measure all joints that occur within its area, or you could draw a line across the outcrop and measure all joints that cross that line (Figure 7.12). The inventory method is necessary if you need to determine fracture density in a body of rock, or provide statistics on joint data. Mathematical procedures for determining joint density in three dimensions from measurements on two dimensional surfaces are available, but are beyond the scope of this book. You can use the inventory method for either systematic or nonsystematic joints.

The inventory method allows you to determine dominant joint orientations using statistical methods. But the problem with the inventory method is that a large number of nonsystematic fractures in an outcrop may obscure the existence of sets of systematic joints, especially if the systematic joints are widely spaced. For this reason, the **selection method** is a more appropriate approach. In this method, you visually scan the outcrop and subjectively decide on the dominant sets (Figure 7.12). Then, you measure a few representative joints of each set and specify the spacing between joints in the set. Effectively, you are filtering your measurements in the field. While this technique won't permit determination of fracture density or provide statistics on joint orientation, it will allow you to define fracture systems in a region. The hazard with this procedure is that a careless observer may record what he or she wants to see, not what is really in the outcrop. When an observer starts scanning a new outcrop, she may subconsciously look for the same joint sets that she had seen in the last outcrop studied, and therefore may miss different, but important, joint sets that occur in the second outcrop, just because they weren't present in the first one.

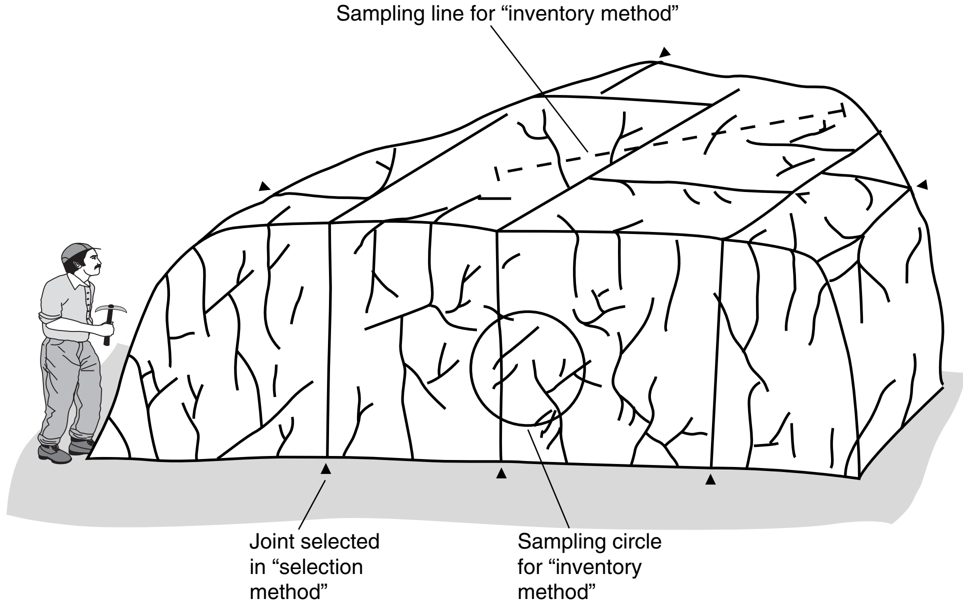


FIGURE 7.12 Joint study using the inventory and selection methods.

Joint data can be recorded in a number of ways. One way is to plot the strike and dip of joints on a geologic map. If the joints are vertical, a measurement of the trend of the joint may be sufficient. You can create a powerful visual impression of joint attitudes in a region, by drawing representative **joint trajectories** as trend lines on a map (Figure 7.13a). These trajectories are lines that represent the trends of joints, but they are not necessarily the map traces of individual joints.

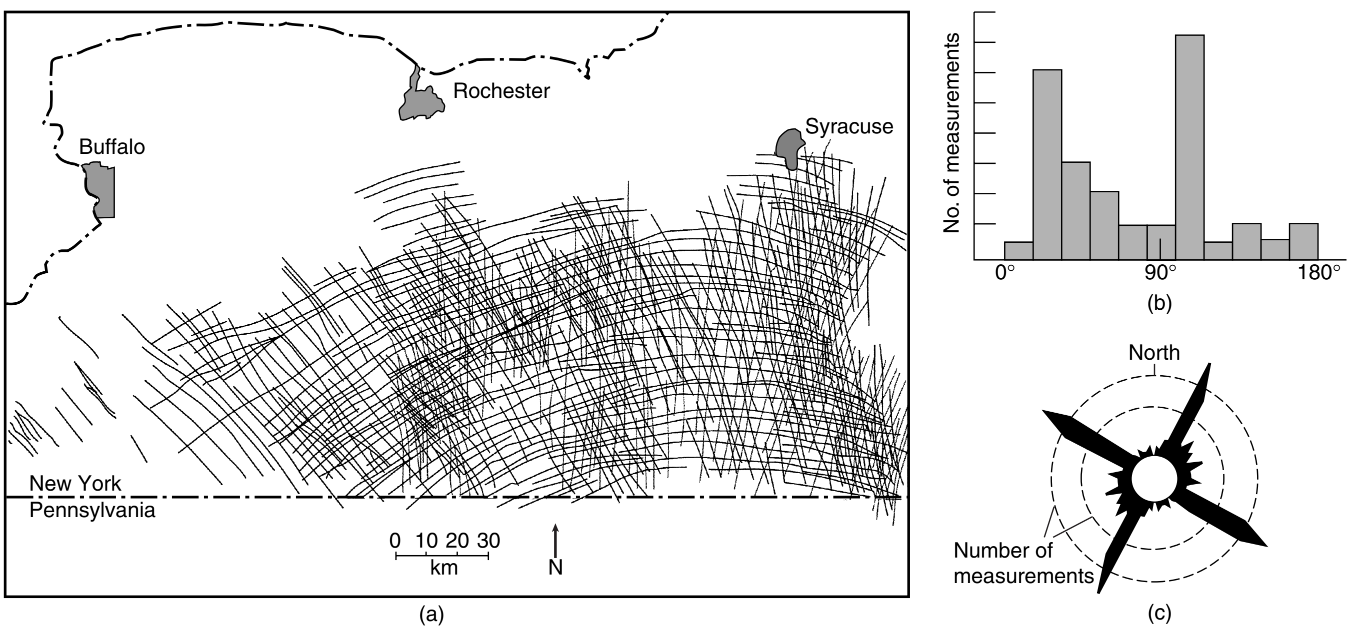


FIGURE 7.13 Ways of representing joint arrays. (a) Joint trajectory map. (b) Frequency diagram (histogram). (c) Rose diagram. The three examples do not portray the same data sets.

Statistical diagrams that show attitudes of many different joints within a given region can help you identify dominant joint orientations in a region. What may appear to be a meaningless jumble at first sight, may resolve into significant groupings upon analysis. If joints in a particular region are not vertical, it is most appropriate to plot their attitudes on a contoured equal-area net, but if the joints are mostly vertical, a common occurrence in flat-lying sedimentary strata, their strikes can be shown on histograms. A **histogram,** in the case of joints, indicates the number of joints whose strike falls within a particular range. On a bar histogram (Figure 7.13b), the abscissa represents bearings from 0° to 180°, and the ordinate is proportional to the number of fracture-strike measurements. On a polar histogram, also called a **rose diagram,** you show the bearings directly on the diagram (Figure 7.13c). The number of joints whose strike falls within a given range is shown by a pie-slice segment whose radius is proportional to the number, or to the percentage, of joints with that orientation. Rose diagrams work better than bar histographs to give you an intuitive feel for the distribution of joint attitudes.

The orientation of joints is not the only information to record during a joint study, as you can see from the list of questions that we presented in the previous section. In modern joint studies, you should also record joint spacing, joint trace length, cross-cutting relations between joints, the relation of joints to lithology, joint surface morphology, and the relation of joints to other structures. To clarify relations among joints, it often helps to add outcrop sketches to your notes.

7.5 ORIGIN AND INTERPRETATION OF JOINTS

Why do joints form? In Chapter 6, we learned that joints develop when stress exceeds the tensile fracture strength of a rock, and Griffith cracks begin to propagate. But under what conditions in the Earth's brittle crust are stresses sufficient to crack rocks? In this section, we describe several possible settings to explain how stress states leading to joint formation develop in a rock body. But before you read further, a note of caution. When using these ideas as a basis for field interpretation of jointing, keep in mind that different joints in the same outcrop may have formed at different times and for different reasons. Once formed, a joint doesn't heal and disappear unless the rock gets metamorphosed or becomes pervasively deformed. Further, local variations in the stress field, which are a natural feature of inhomogeneous rock, may cause joints that formed at the same time to have different orientations at different locations. Because of these factors, joint interpretation continues to challenge geologists, and will do so for years to come.

7.5.1 Joints Related to Uplift and Unroofing

Lithostatic pressure due to the weight of overlying rock compresses rock at depth. Also, because of the Earth's geothermal gradient, rock at depth is warmer than rock closer to the surface. Subsequent regional uplift leads to erosion of the overburden and the unroofing of buried rock (Figure 7.14). This unroofing causes a change in the stress state for three reasons: cooling, the Poisson effect, and the membrane effect.

As the burial depth of rocks decreases, they *cool and contract.* The rock can shrink in a vertical direction without difficulty, because the Earth's surface is a free surface. But, because the rock is embedded in the earth, it is not free to shrink elastically in the horizontal direction as much as if it were unconfined, so horizontal tensile stress develops in the rock. Furthermore, as the overburden diminishes, rock expands (very slightly) in the vertical direction. Therefore, because of the **Poisson effect** (see Chapter 6), it contracts in the

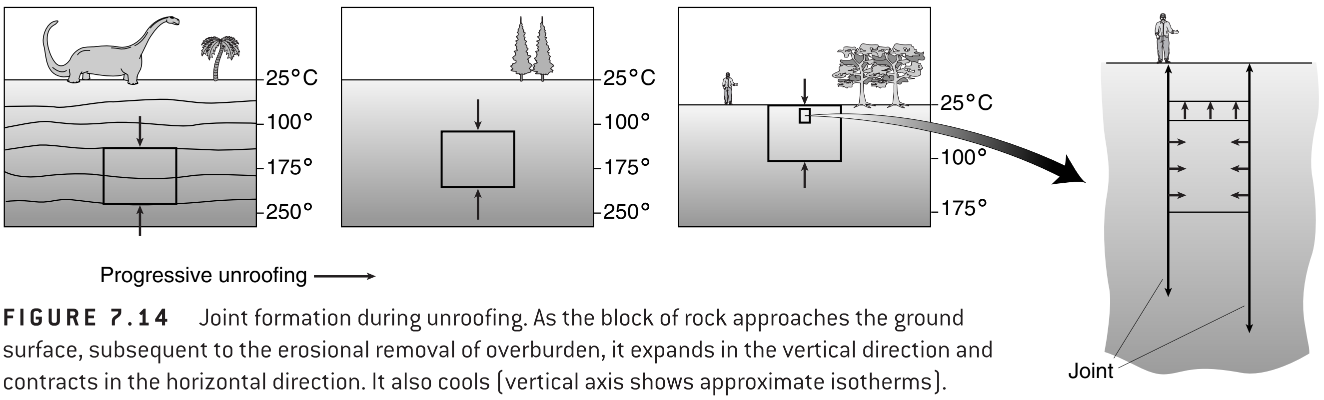


FIGURE 7.14 Joint formation during unroofing. As the block of rock approaches the ground surface, subsequent to the erosional removal of overburden, it expands in the vertical direction and contracts in the horizontal direction. It also cools (vertical axis shows approximate isotherms).

horizontal direction. Again, because the rock is embedded in the earth, it cannot shorten in the horizontal direction as much as it would if it were unconfined, so a horizontal tensional stress develops. Uplift and unroofing effectively cause rock layers to move away from the center of Earth. The layer stretches like a membrane as its radius of curvature increases, thereby creating tensile stress in the layer, called the **membrane effect.**

If the horizontal tensional stress created by any or all of these factors overcomes the compressive stresses due to burial and exceeds the rock's tensile strength, it will cause the rock to crack and to form joints. Joints formed by the processes just described tend to be vertical because they generate a horizontal σ_3. Recall that the Earth's surface is a free surface, so it must be a principal plane of stress. Therefore, the other two principal planes of stress must be vertical. Uplift and unroofing are particularly important causes of joint formation in sedimentary basins of continental interiors, which are subjected to epeirogenic movements, and in orogens that are uplifted long after collisional or convergent tectonism has ceased.

7.5.2 Formation of Sheeting Joints

Uplift and exhumation of rocks may lead to the development of sheeting joints within a few hundred meters of the Earth's surface. As we mentioned earlier, sheeting joints are commonly subparallel to topographic surfaces, and are most prominent in rocks that do not contain bedding or schistosity, particularly granitic rocks.

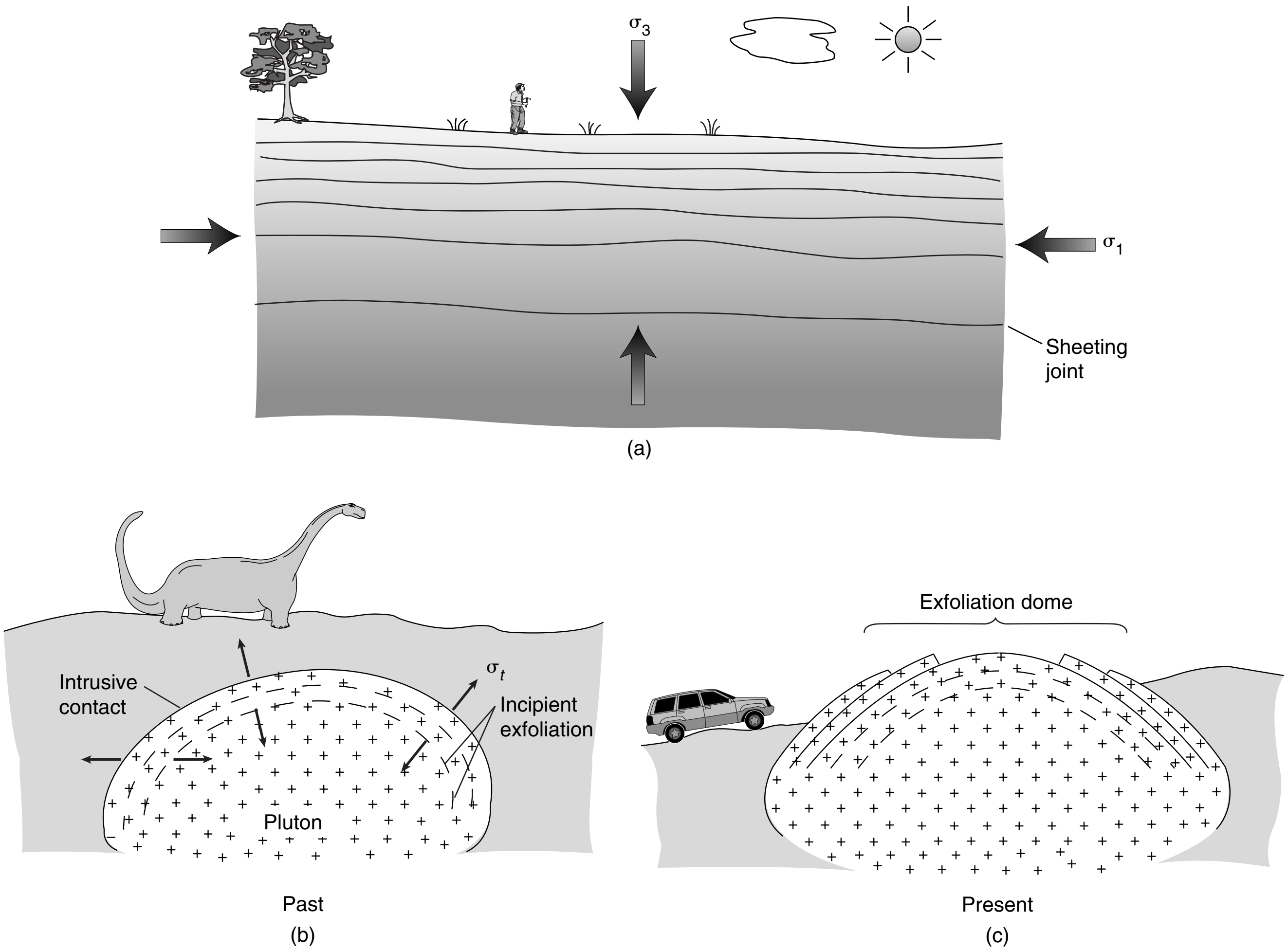


FIGURE 7.15 (a) Sheeting joints forming in a location where σ_1 is horizontal while σ_3 is vertical, near the ground surface. Note that the joints become more closely spaced closer to the ground surface. (b) Consider a situation where a pluton cools and contracts more than country rock, so σ_t (tensile stress) is oriented perpendicular to the intrusive contact. (c) Later, when the pluton is exhumed, joints form parallel to the intrusive contact and create an exfoliation dome.

The origin of sheeting joints is a bit problematic. At first glance, you might not expect joints to form parallel to the ground surface, because they are tensile fractures, and near the ground surface there is a compressive load perpendicular to the ground surface due to the weight of the overlying rock, and the lack of high fluid pressure. It appears that sheeting joints form where horizontal stress is significantly greater than the vertical load (Figure 7.15a), allowing joints to propagate parallel to the ground surface. In this regard, formation of sheeting joints resembles cracks formed by longitudinal splitting in laboratory specimens.

The stresses causing sheeting joints may, in part, be tectonic in origin, but they may also be residual stresses. A **residual stress** remains in a rock even if it is no longer loaded externally (e.g., in an unconfined block of rock sitting on a table). Residual stresses develop in a number of ways. Imagine a layer of dry sand that gets deeply buried. Because of the weight of the overburden, the sand grains squeeze together and strain elastically. If, at a later time, groundwater fills the pores between the strained grains, unstrained cement may precipitate and lock the grains together. As a consequence, the elastic strain in the grains gets locked into the resulting sandstone. When unroofing later exposes the sandstone, the grains and the cement expand by different amounts, and as a consequence stress develops in the sandstone. In the case of pluton, residual stresses develop because its thermal properties (e.g., coefficient of thermal expansion) differ from those of the surrounding wall rock, and because, during cooling, the pluton cools by a greater amount than the wall rock. The pluton and the wall rock tend to undergo different elastic strains as a result of thermal changes during cooling and later unroofing (Figure 7.15b and c). Because the pluton is welded to the surrounding country rock, the differential strain creates an elastic stress in the rock. For example, if the pluton shrinks more than the wall rock, tensile stresses develop perpendicular to the wall. At depth, compressive stress due to the overburden counters these tensile stresses, but near the surface, residual tensile stress perpendicular to the walls of the pluton may exceed the weight of the overburden and produce sheeting joints parallel to the wall of the pluton.

We earlier noted that sheeting joints tend to parallel topography. This relationship either reflects topographic control of the geometry of joints (because the vertical load is perpendicular to the ground surface), or joint control of the shape of the land surface (because rocks spall off the mountainside at the joint surface). Geologists are not sure which phenomenon is more important.

7.5.3 Natural Hydraulic Fracturing

As we saw in Chapter 3, the three principal stresses at depth in most of the continental lithosphere are compressive. Yet joints form in these regions, and these joints may be decorated with plumose structure indicating that they were driven by tensile stress. How can joints form if all three principal stresses are compressive? As we described in Chapter 6, the solution to this paradox comes from considering the effect of pore pressure on fracturing. In simple terms, the increase in pore pressure in a preexisting crack pushes outward and causes a tensile stress to develop at the crack tip that eventually exceeds the magnitude of the least principal compressive stress. If the pore pressure is sufficiently large, a tensile stress that exceeds the magnitude of σ_3 develops at the tip the crack, even if the remote principal stresses are all compressive (Figure 7.16a), and the crack propagates, a process called **hydraulic fracturing.** Oil well engineers commonly use hydraulic fracturing to create fractures and enhance permeability in the rock surrounding an oil well. They create hydraulic fractures by increasing the fluid pressure in a sealed segment of the well until the wall rock breaks. But hydraulic fracturing also occurs in nature, due to the fluid pressure of water, oil, and gas in rock, and it is this *natural* hydraulic fracturing that causes some joints to form.

If you think hard about the explanation of hydraulic fracturing that we just provided, you may wonder whether the process implies that the pore pressure in the crack becomes greater than pore pressure in the pores of the surrounding rock. It doesn't! Pore pressure in the crack can be the same as in the pores of the surrounding rock during natural hydraulic fracturing. Thus, we need to look a little more closely at the problem to understand why pore pressure can cause joint propagation.

Imagine that a cemented sandstone contains fluid-filled pores and fluid-filled cracks (Figure 7.16b). Let's focus our attention on the crack and its walls. Because the pores and the crack are connected, the fluid pressure in the pores and the crack are the same. Fluid pressure within the crack is pushing outwards, creating an opening stress, but at the same time, the fluid pressure in the pores, as well as the stress in the rock, is pushing inwards, creating a closing stress. As long as the closing stress exceeds the opening stress, the crack does not propagate. If the fluid pressure increases, the opening stress increases at the same rate as the increase in fluid pressure, but the closing stress increases at a slower rate. Eventually, the open-

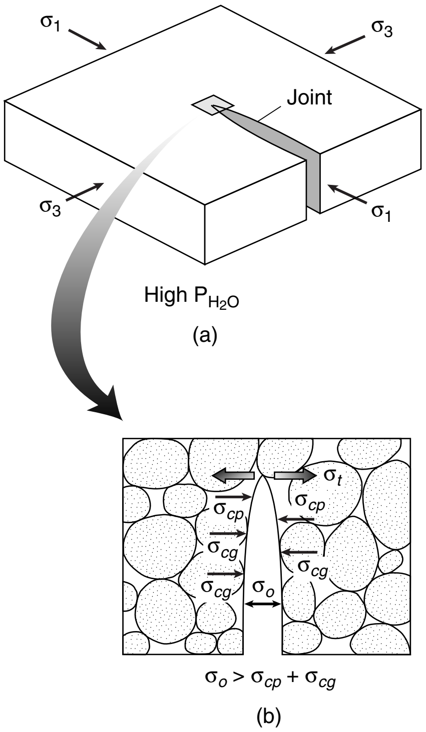


FIGURE 7.16 (a) Block diagram showing the stresses in the vicinity of a crack in which there is fluid pressure that exceeds the magnitude of σ_3. As a result, there is a tensile stress, σ_t, along the crack. (b) Enlargement of the crack tip, illustrating the poroelastic effect. The opening stress (σ_o) due to fluid pressure in the crack exceeds the closing stress (σ_c), which is the sum of σ_{cp}, the closing stress where a pore is in contact with the crack, and σ_{cg}, the closing stress where a grain is in contact with the crack.

ing stress exceeds the closing stress so that the crack propagates; effectively, the outward push of the fluid in the crack creates a tensile stress at the crack tip. Why does the closing stress increase at a slower rate than the fluid pressure and the opening stress? The reason is that grains in the rock are cemented to one another, so that the grains cannot move freely in response to the increase in fluid pressure in the pores. The elasticity of the grains themselves, therefore, takes up some of the push caused by the fluid pressure. Thus, the closing stress acting on the fluid in the crack where it is in contact with a grain is less than where the fluid in the crack is in contact with a pore, but the outward push of the fluid in the crack is the same everywhere.[5] As a result, the net outward push exceeds the net inward push, and tensile stress locally develops.

Once the crack propagates, the volume of open space between the walls of the crack increases, so the fluid pressure in the crack decreases. As a consequence, the crack stops growing until an increase in fluid pressure once again allows the stress intensity at the crack tip to drive the tip into unfractured rock. Thus, the surfaces of joints formed by natural hydraulic fracturing tend to have many arrest lines.

7.5.4 Joints Related to Regional Deformation

During a convergent or collisional orogenic event, compressive tectonic stress affects rocks over a broad region, including the continental interior. Joints form within the foreland of orogens during tectonism for a number of reasons.

Joints from natural hydrofracturing often form on the foreland margins of orogens during orogeny. The conclusion that the joints are syntectonic is based on two observations. First, the joints parallel the σ_1 direction associated with the development of tectonic structures like folds. Second, the joints locally contain mineral fill which formed at temperatures and fluid pressures found at a depth of several kilometers; thus, they are not a consequence of the recent cracking of rocks in the near surface. The origin of such joints may reflect increases in fluid pressure within confined rock layers due to the increase in overburden resulting from thrust-sheet emplacement, or from the deposition of sediment eroded from the interior of the orogen.

During an orogenic event, the maximum horizontal stress is approximately perpendicular to the trend of the orogen. As a consequence, the joints that form by syntectonic natural hydraulic fracturing are roughly perpendicular to the trend of the orogen. Because the stress state may change with time in an orogen, later-formed joints may have a different strike than earlier-formed joints, and the joints formed during a given event might not be exactly perpendicular to the fold trends where they form. Such joint patterns are typical of orogenic foreland regions, but may also occur in continental interiors.

Joints are commonly related to faulting, and these fall into three basic classes. The first class is composed of regional joints that develop in the country rock due to the stress field that is also responsible for generation and/or movement on the fault itself. Since faults are usually inclined to the remote σ_1 direction, the joints that form in the stress field that cause a fault to move

[5] This is known as the *poroelastic effect.*

will not be parallel to the fault (Figure 7.17a). The second class includes joints that develop due to the distortion of a moving fault block. For example, the hanging wall of a normal fault bock may undergo some extension, resulting in the development of joints, or the hanging block of a thrust fault may be warped as it moves over the fault, if the underlying fault surface is not planar, and thus may locally develop tensile stresses sufficient to crack the rock (Figure 7.17b). The third class includes joints that form immediately adjacent to a fault in response to tensile stresses created in the wall rock while the fault moves. Specifically, during the development of a shear rupture (i.e., a fault), an *en echelon* array of short joints forms in the rock adjacent to the rupture. These joints merge with the fault and are inclined at an angle of around 30° to 45° to the fault surface; they are called **pinnate joints** (Figure 7.17c).

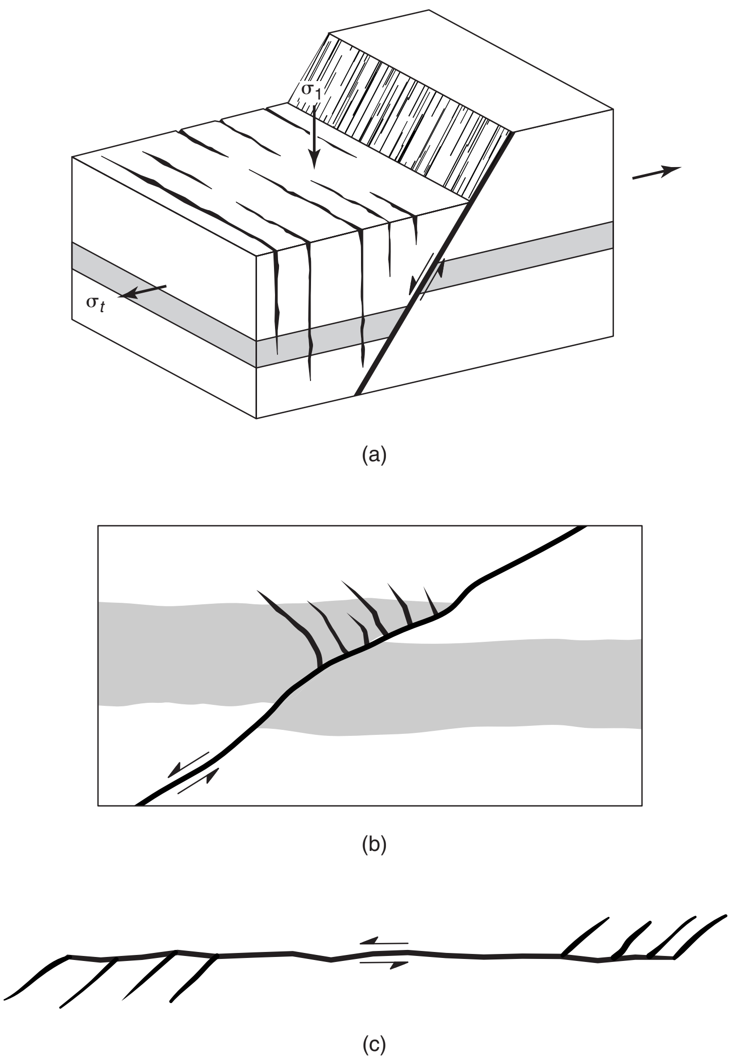


FIGURE 7.17 (a) Formation of joints in the hanging-wall block of a region in which normal faulting is taking place. (b) Formation of joints above an irregularity in a (reverse) fault surface. (c) Pinnate joints along a fault.

The acute angle between a pinnate joint and the fault indicates the sense of shear on the fault, as we will see in Chapter 8.

When the stress acting on a region of crust is released, the crust elastically relaxes to attain a different shape. This change in shape may create tensile stresses within the region that are sufficient to create **release joints,** such as occurs in relation to folding. Folded rocks may be cut by syntectonic natural hydrofractures, a process manifested by joints oriented at a high angle to the fold-hinge, as we described previously (Figure 7.18). In addition, during the development of folds in nonmetamorphic conditions, joints often develop because of local tensile stresses associated with bending of the layers (Figure 7.18). Joints resulting from this process of outer-arc extension have a strike that is parallel to the trend of the fold-hinge, and may converge toward the core of the fold. If development of folds results in stretching of the rock layer parallel to the hinge of the fold, then cross-strike joints may develop.

Finally, joints may develop in a region of crust that has been subjected to broad regional warping, perhaps due to **flexural loading** of the crust. Like folding, joint formation reflects tensile stresses that develop when the radius of curvature of a rock layer changes.

7.5.5 Orthogonal Joint Systems

In orogenic forelands and in continental interiors, you will commonly find two systematic joint sets that are mutually perpendicular. In some cases, the joints define a **ladder pattern** (Figure 7.19a), in which the joints of one set are relatively long, whereas the joints of the other are relatively short cross joints which terminate at the long joints. In other cases, the joints define a **grid pattern** (Figure 7.19b), in which the two sets appear to be mutually cross cutting. The existence of such orthogonal systems has perplexed geologists for decades, because at first glance it seems impossible for two sets of tensile fractures to form at 90° to one another in the same regional stress field. Recent field and laboratory studies suggest a number of possible ways in which orthogonal systems develop, though their application to specific regions remains controversial.

In orogenic forelands, an orthogonal joint system typically consists of a strike-parallel and a cross-strike set, defining a ladder pattern. The two sets may have quite different origins.

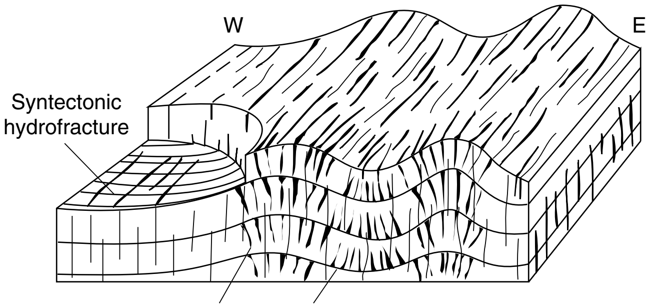


FIGURE 7.18 Block diagram showing outer-arc extension joints whose strike is parallel to the hinge of folds.

Cross-strike joints parallel the regional maximum horizontal stress trajectory associated with folding, and thus may have formed as syntectonic natural hydrofractures, whereas strike-parallel joints could reflect outer-arc extension of folded layers. Alternatively, the strike-parallel joints could be release joints formed when orogenic stresses relaxed.

Orthogonal joint systems may develop in regions that were subjected to a regional tensile stress that was later relaxed. During the initial stretching of the region, a set of joints develops perpendicular to the regional tensile stress. When the stress is released, the region rebounds elastically, and expands slightly in the direction perpendicular to the original stretching. A new set of joints therefore develops perpendicular to the first.

Orthogonal joint systems may also develop during uplift. Imagine that a rock layer is unloaded when the overburden erodes away. As a result, a joint set perpendicular to the regional σ_3 develops. With continued uplift and expansion, the tensile stress that develops in the layer can be relieved easily in the direction perpendicular to the existing joints; that is, they just open up. But tensile stresses cannot be relieved in the direction parallel to the existing joints, so new cross joints form, creating a ladder pattern.

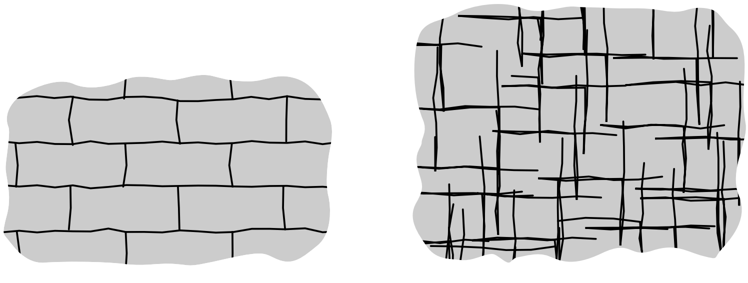


FIGURE 7.19 Two patterns of orthogonal joint systems. (a) Traces of joints defining a ladder pattern. (b) Traces of joints defining a grid pattern.

Grid patterns (Figure 7.19b) suggest that the two joint sets initiated at roughly the same time, or that cracking episodes alternated between forming members of one set and then the other. If we assume that both joint sets form in the principal plane that is perpendicular to σ_3, we can interpret such occurrences as being related to the back-and-forth interchange of σ_2 and σ_3 during uplift, when σ_2 and σ_3 are similar in magnitude. To see what we mean, imagine a region where σ_1 is vertical, and σ_3 is initially north–south. When σ_3 is north–south, east–west trending joints develop. But if σ_3 switches with σ_2 and becomes east–west, then north–south trending joints develop.

7.5.6 Conjugate Joint Systems

At some localities in orogenic forelands we find that joint sets define a conjugate system in which the bisector of the dihedral angle is perpendicular to the axis of folds. The origin of such fracture systems remains one of the most controversial aspects of joint interpretation. Based on their geometry, it was traditionally assumed that conjugate joints are either shear fractures, formed at about 30° to σ_1 (representing failure when the Mohr circle touches the Coulomb failure envelope; Chapter 8), or so-called transitional-tensile fractures that are thought to form at angles less than 30° to σ_1 (representing failure when the Mohr circle touches the steep part of the failure envelope; see Chapter 8). Yet, if you examine the surfaces of the joints in these conjugate systems, you find in many cases that they display plumose structure, confirming that they formed as Mode I (extension) fractures. Further, as we noted earlier, transitional-tensile fractures have never been created in the lab, so their very existence remains suspect. The only type of crack that is known to propagate for long distances in its own plane is a Mode I crack; shear fractures form by linkage of microcracks, not by propagation of a single shear surface in its own plane. But if the members of conjugate joint systems are not shear fractures, how do they form?

Many researchers now believe that both of the two nonparallel sets in the conjugate system are cross-strike joints that initially formed perpendicular to σ_3. Thus, to explain the contrast in orientation between the two sets, they suggest that the two sets formed at different times in response to different stress fields. For example, the slightly folded Devonian strata of the Appalachian Plateau in south-central New York State contain two joint sets separated by an angle of about 60° (Figure 7.13a).

FIGURE 7.20 Large joint face in Entrada sandstone near Moab (Utah, USA). Note how the cliff face is a large joint surface. The thin bedded shale unit below the sandstone has more closely spaced joints.

The two sets are attributed to different, distinct phases of the Late Paleozoic Alleghanian orogeny; the maximum horizontal stress during the first phase of the orogeny was not parallel to that during the second phase of orogeny. Geologists who favor this model for conjugate joint systems conclude that the occurrence of slip lineations on joints of conjugate systems does not imply that the joints originated as shear fractures, but rather that they reactivated as mesoscopic faults subsequent to their formation.

7.5.7 Joint Trend as Paleostress Trajectory

Orientation data on jointing holds valuable information about the orientation of stress fields at the time of failure. Because joints propagate normal to σ_3, their planes define the trajectories of σ_3 in a region. In the case of vertical joints, the strike of the joint defines the trajectory of maximum horizontal stress (σ_H). We don't a priori know if σ_H represents σ_1 or σ_2. The maximum principal stress could be either parallel or perpendicular to the Earth's surface, depending on the depth at which the joint formed.

7.6 LIMITS ON JOINT GROWTH

Having discussed various ways in which joints initiate, we also need to address the issue of why joints stop growing. Recall from Chapter 6 that the stress intensity at the tip of a crack depends on the length of the crack. Thus, the stress intensity increases as the crack grows, and as long as the stress driving joint growth remains unchanged, the joint will keep growing. For this reason, joints that grow in large bodies of homogeneous rock can become huge surfaces, as seen in the massive beds of sandstone in southern Utah shown in Figure 7.20. But joints clearly do not propagate from one side of a continent to the other. They stop growing for one or more of the following reasons.

The joint tip may *intersect a (nearly) free surface.* Joints obviously stop growing when they reach Earth's surface. They stop growing in the subsurface where they intersect a preexisting open fracture (joint or fault) or a weak bedding plane, or where they pass downward into ductile rock. Two joints that are growing towards each other, but are not coplanar, stop growing when they enter each other's stress shadow (Figure 7.21a). In some cases, the *interaction of joint tips* causes curvature where the two joints link (Figure 7.21b). If, however, the preexisting joint is squeezed together so tightly that friction allows shear stress to be transmitted across it, or if it has been sealed by vein material, then a younger joint can cut across it.

Formation of the joint itself may cause a local *drop in fluid pressure,* because creation of the joint creates space for fluid. This increase in space temporarily causes a drop in fluid pressure, so that the stress intensity at the joint tip becomes insufficient to propagate into unfractured rock. When fluid flow into the joint increases the fluid pressure to a large enough value, the joint growth resumes. Thus, as mentioned earlier, it is

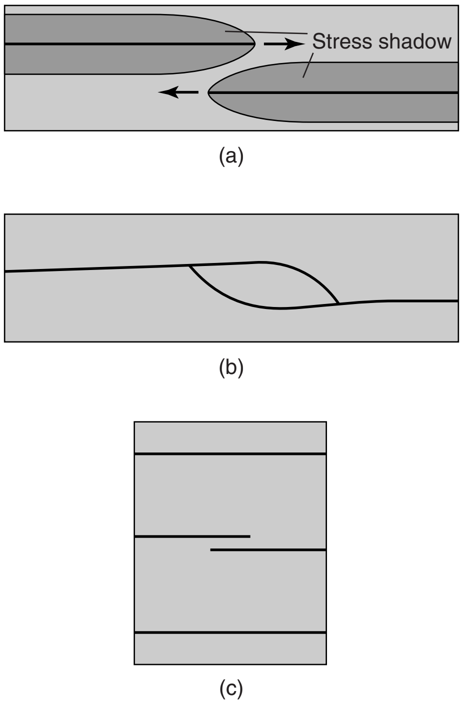


FIGURE 7.21 Joint terminations. (a) Joints terminating without curving when they approach one another. (b) Joints curving into each other and linking. (c) Map view sketch illustrating how joint spacing is fairly constant because joints that grow too close together cannot pass each other.

characteristic of joints being driven by high fluid pressures to grow in a start-stop manner, so their surfaces show many arrest lines.

Finally, if the joint grows into a region where energy at the crack tip can be dissipated by plastic yielding, the joint stops growing. Similarly, propagation of a joint into a rock with a *different stiffness* or tensile strength may cause it to stop growing. Also, if the joint tip enters a region where the stress intensity at the crack tip becomes too small to drive the cracking process, then the joint stops growing. The decrease in stress intensity may be due to a decrease in the tensile stress magnitude in the rock, or due to an increase in compressive stress that holds the joint together.

7.7 VEINS AND VEIN ARRAYS

In the vast desert ranges of western Arizona, there are few permanent residents, save for the snakes and scorpions, but almost every square meter of the rugged terrain has been trod upon by a dusty prospector in search of valuable deposits of gold, silver, or copper. Modern geologists mapping in the region frequently come upon traces of prospectors from years past. When you poke into many of these excavations, you find that the focus of their efforts, the days and days of agonizing labor with pick and shovel, is nothing more than a vein of milky white quartz. What are veins? Simply speaking, a **vein** is a fracture filled with mineral crystals that precipitated from a watery solution (Figure 7.1b; Table 7.2). Quartz or calcite form the most common vein fill, but other minerals do occur in veins, including numerous ore minerals, zeolites, and chlorite. Some veins initiated as joints, whereas others initiated as faults or as cracks adjacent to faults. Veins come in all dimensions; some are narrower and shorter than a strand of hair, while others comprise massive tabular accumulations that are meters across and tens of meters long. Groups of veins are called **vein arrays** and these have a variety of forms, as described in Table 7.2.

TABLE 7.2	VEIN TERMINOLOGY
Vein	A fracture that has filled with minerals precipitated from water solutions that passed through the fracture.
Vein array	A group of veins in a body of rock.
Planar systematic arrays	In a planar systematic array, the component veins are planar, are mutually parallel, and are regularly spaced.
Nonsystematic arrays	The veins in nonsystematic arrays tend to be nonplanar, and individual veins may vary in width.
En echelon arrays	An *en echelon* array consists of short parallel veins that lie between two parallel enveloping surfaces and are inclined at an angle to the surfaces. Typically, the veins in an *en echelon* array taper toward their terminations. The veins may be sigmoidal in cross section.
Stockwork veins	Stockwork veins comprise a cluster of irregularly shaped veins that occur in a pervasively fractured rock body. The veins are nonplanar arrays and occur in a range of orientations. In rock bodies with stockwork veins, as much as 40% or 50% of the outcrop is composed of vein material, and vein material may completely surround blocks of the host rock.

(a) (b)

FIGURE 7.22 Vein arrays. (a) Planar array of veins. (b) Stockwork array of veins. Vein fill is dark.

7.7.1 Formation of Vein Arrays

Planar systematic arrays (Figure 7.22a) represent mineralization of a preexisting systematic joint set or mineralization during formation of a systematic joint set. **Stockwork vein arrays** (Figure 7.22b) form where rock has been shattered, either by the existence of locally very high fluid pressure, or as a result of pervasive fracturing in association with folding and faulting.

En echelon vein arrays form in a couple of different ways. They may form by filling *en echelon* joints in the twist hackle fringe of a larger joint. As we saw earlier in this chapter, the twist hackle fringe represents the breakup of a joint into short segments when it enters a region of the rock with a different stress field. *En echelon* vein arrays also develop as a consequence of shear within a rock body that is associated with displacement across a fault zone (Figure 7.23). The fractures comprising an *en echelon* array initiate parallel to σ_1, typically at an angle of about 45° to the borders of the shear. Fractures open as displacement across the shear zone develops, and fill with vein material (Figure 7.23b). Once formed, the veins are material objects within the rock, so continued shear will rotate the veins and the angle increases. If, however, a new increment of vein growth occurs at the tip, these new increments initiate at 45° to the shear surfaces. Therefore, the veins become sigmoidal in shape (Figure 7.23c). Locally, a second generation of veins may initiate at the center of the original veins; this second set cuts obliquely across the first generation of veins. Because of the geometric relationship between *en echelon* veins and displacement, the orientation of shear-related *en echelon* veins can be used to determine shear sense (see Chapters 8 and 12).

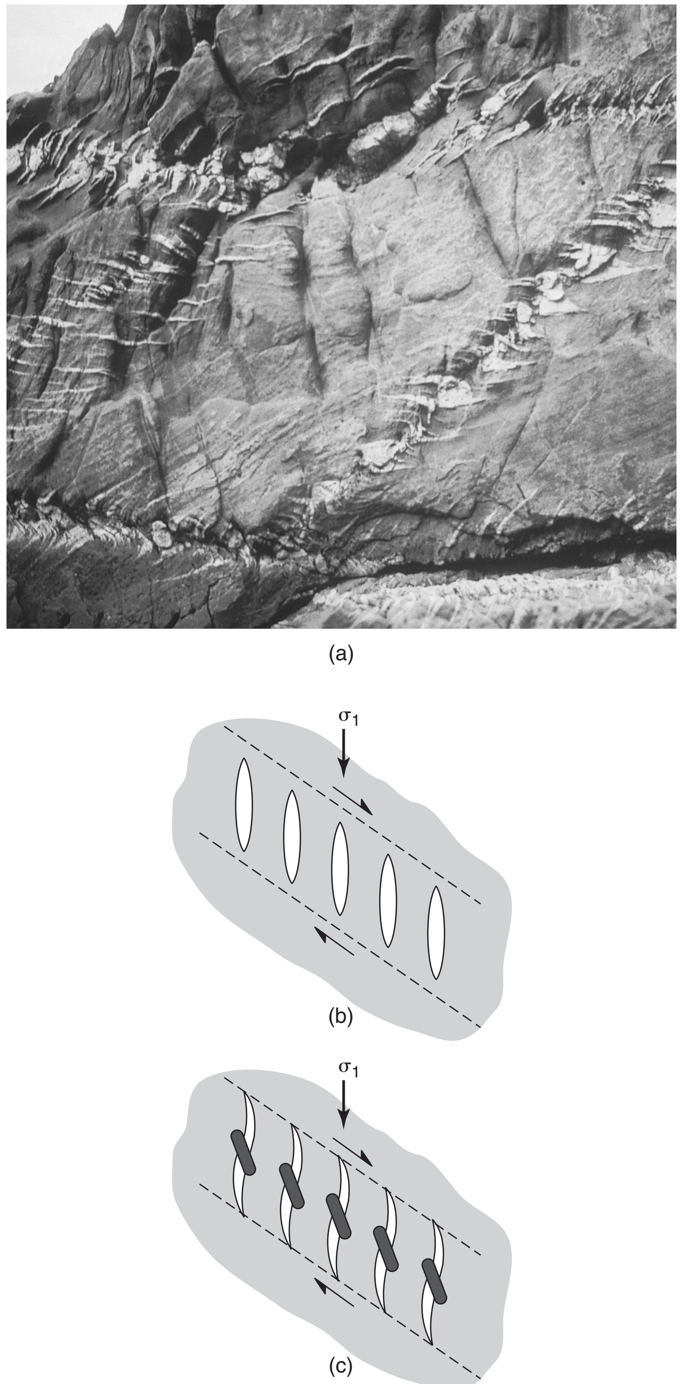


FIGURE 7.23 (a) *En echelon* veins in the Lachlan Orogen (southeastern Australia). (b) Formation of a simple *en echelon* array. (c) Formation of sigmoidal *en echelon* veins, due to rotation of the older, central part of the veins, and the growth of new vein material at ~45° to the shear surface.

7.7.2 Vein Fill: Blocky and Fibrous Veins

Vein fill, the mineral crystals within a vein, is either blocky (also called sparry) or fibrous. In **blocky veins,** the crystals of vein fill are roughly equant, and may exhibit crystal faces (Figure 7.24a). The occurrence of blocky veins means that the vein was an open cavity when the mineral precipitated (this is possible only in veins formed near the surface, where rock strength is sufficient to permit a cavity to stay open or fluid pressure is great enough to hold the fracture open), that

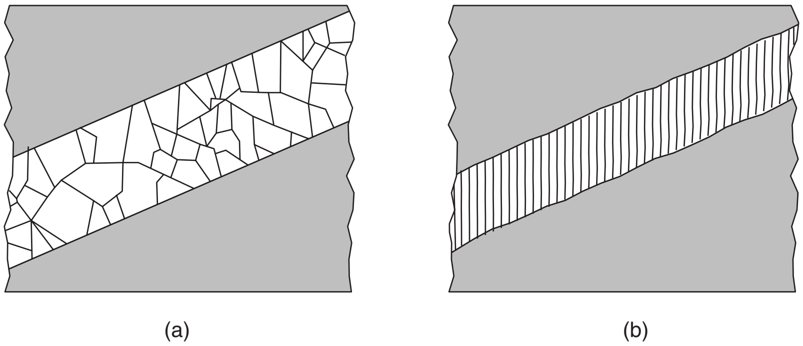


FIGURE 7.24 Vein fill types. (a) Blocky vein fill. (b) Fibrous vein fill.

previously formed vein fill later recrystallized to form blocky crystals, or that there were few nucleation sites for crystals to grow from during vein formation.

In **fibrous veins,** the crystals are long relative to their width, so that the vein has the appearance of being spanned by a bunch of hairs (Figure 7.24b). Geologists don't fully agree on the origin of fibrous veins, but some fibrous veins may form by the **crack-seal process.** The starting condition for this process is an intact rock containing pore fluid that in turn contains dissolved minerals. If the fluid pressure becomes great enough, the vein cracks and a very slight opening (only microns wide) develops. This crack immediately fills with fluid; but since the fluid pressure within the open crack is less than in the pores of the surrounding rock, the solubility of the dissolved material decreases and the mineral precipitates, thereby sealing the crack. The process repeats tens to hundreds of times, and each time the vein width grows slightly (Figure 7.25). During each increment of growth, existing grains in the vein act as nuclei on which the new vein material grows, and thus continuous crystals grow. Figure 7.25 shows microscopic evidence for this crack-seal process in the formation of a fibrous vein. Alternatively, formation of some fibrous veins may occur by a diffusion process, whereby ions migrate through fluid

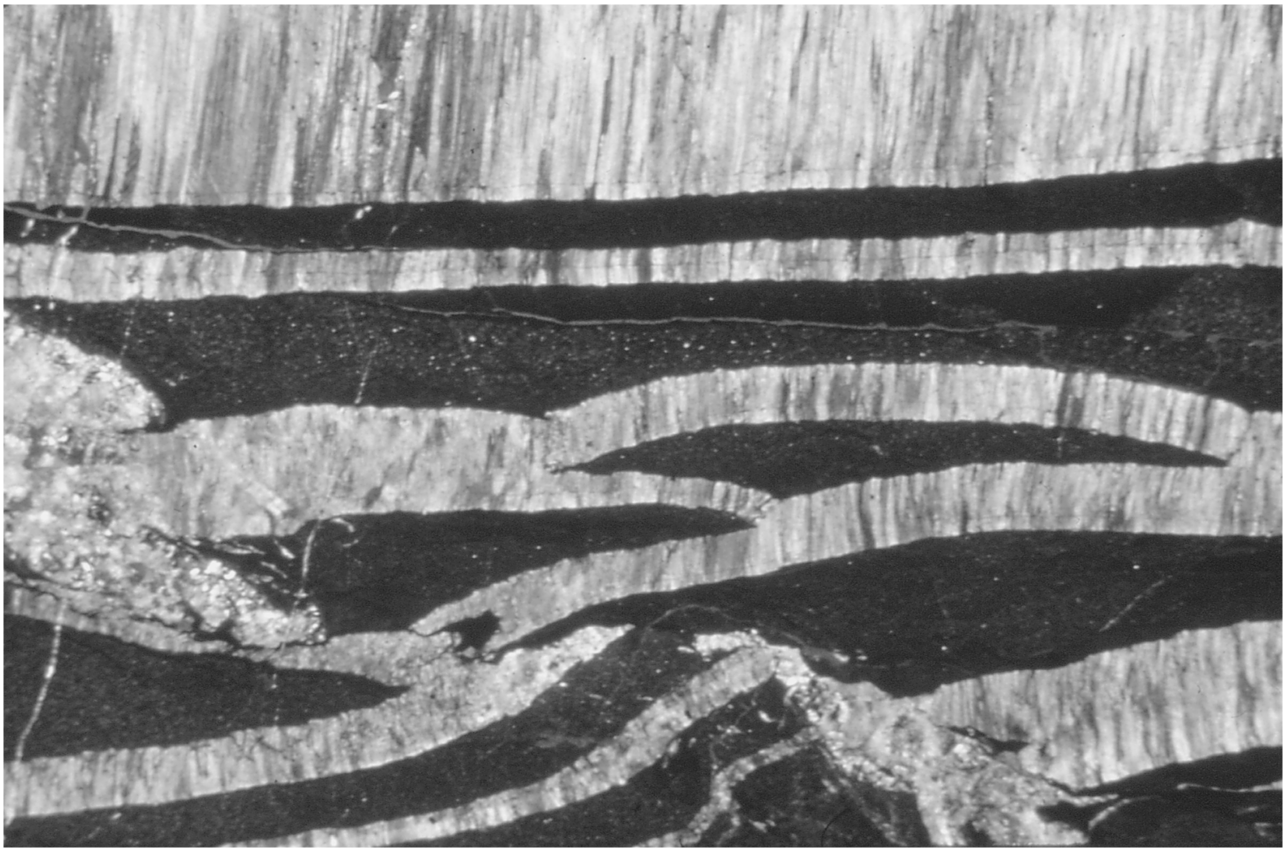


FIGURE 7.25 Photomicrograph of fibrous calcite filling in tensile fractures. Width of view is ~3 cm.

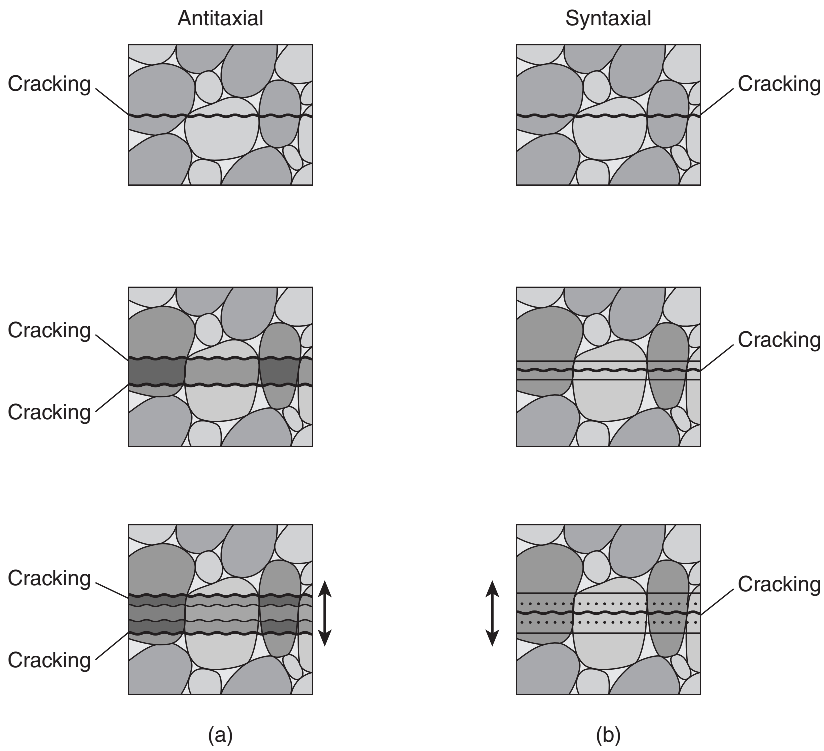


FIGURE 7.26 Cross-sectional sketches, at the scale of individual grains, showing the contrast in the stages of crack-seal deformation leading to antitaxial veins and syntaxial veins. (a) Formation of antitaxial veins. The increments of cracking form along the margins of the vein, and the vein composition differs from the wall rock (i.e., the fibers are not in optical continuity with the grains of the wall). During increments of cracking, tiny slices of the wall rock spall off. The slices bound the growth increments in a fiber. (b) Syntaxial vein formation. During each increment, the cracking is in the center of the vein. The composition of the fibers is the same as that of the grains in the wall rock (that is, the fibers are in optical continuity with the grains of the wall rock). Optical continuity between fiber and grain means that the crystal lattice of the grain has the same orientation as the crystal fiber of the fiber. Optically continuous fibers and grains go extinct at the same time, when viewed with a petrographic microscope.

films on grain boundaries and precipitate at the tips of fibers while the vein walls gradually move apart. During this process, an open crack never actually develops along the vein walls or in the vein.

7.7.3 Interpretation of Fibrous Veins

Fibrous veins, in particular those consisting of calcite and quartz, are interesting because they can record useful information about the progressive strain history in an outcrop. There are two end-member types of fibrous veins, syntaxial and antitaxial veins (Figure 7.26).

Syntaxial veins typically form in rocks where the vein fill is the same composition as the wall rock; for example, quartz veins in a quartz sandstone. The vein fibers nucleate on the surface of grains in the wall rock and grow inwards to meet at a median line. Each successive increment of cracking occurs at the median line, because at this locality separate fibers meet, whereas at the walls of the vein, vein fibers and grains of the wall rock form single continuous crystals. Each growth increment of a fiber is bounded by a trail of fluid inclusion.

Antitaxial veins form in rocks where the vein fill is different from the composition of the wall rock; for example, a calcite vein in a quartz sandstone. In antitaxial veins, the increments of cracking occur at the boundaries between the fibers and the vein wall, probably because that is where the bonds are weakest. Thus, antitaxial veins grow outward from the center. Increments of growth are sometimes bounded by trails of small dislodged flakes of the wall rock.

In many cases, the long axis of a fiber in a fibrous vein tracks the direction of maximum extension (stretching) at the time of growth (i.e., it parallels the long axis of the incremental strain ellipsoid). When fibers are perpendicular to the walls of the vein, the vein progressively opened in a direction roughly perpendicular to its walls (Figure 7.27a). However, vein fibers oblique to the vein walls indicate that the vein opened obliquely and that there was a component of shear displacement during vein formation (Figure 7.27b). When vein fibers are sigmoidal in shape, the extension direction rotated relative to the vein-wall orientation. Note that for identical-looking fibers, the order of the movement stages depends on whether the vein is antitaxial or syntaxial (Figure 7.27c-d). For example, if the fibers in a syntaxial vein are perpendicular to the walls in the center and oblique to the walls along the margins, it means that in the early stage of vein formation the vein had an oblique opening component, while at a later stage it did not (remember that the fibers grow toward the center). However, fibers with exactly the same shape in an antitaxial vein would indicate the opposite strain history, because in antitaxial veins the fibers grow toward the walls. The

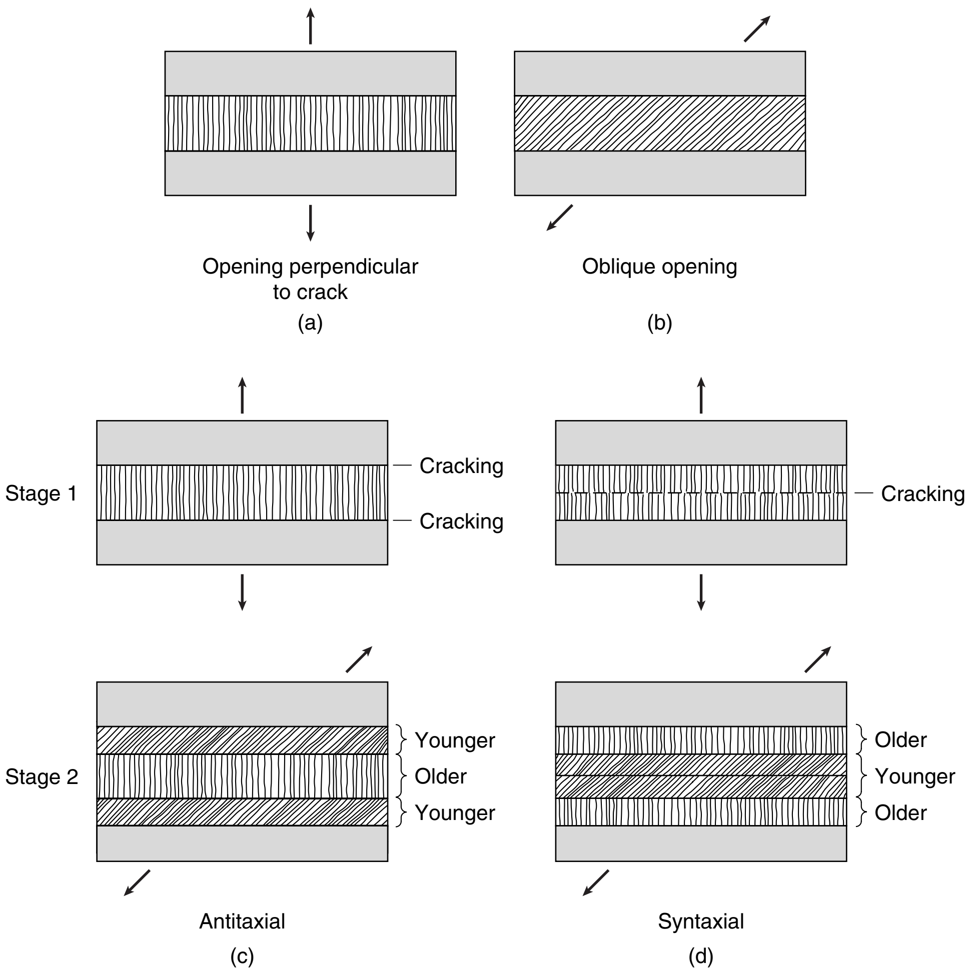


FIGURE 7.27 Cross-sectional sketches showing that the long axis of fibers in a vein tracks the direction of extension, and how a change in extension direction leads to the formation of sigmoidal fibers. (a) If the extension direction is perpendicular to the vein wall then the fibers are perpendicular to the grain. (b) If the extension direction is oblique to the vein wall, then the fibers are oblique to the vein wall. If the extension direction is effectively parallel to the vein surface (i.e., the vein is a fault surface), then the fibers are almost parallel to the vein wall, and on exposed surface would form slip lineations. (c, d) A change in extension direction leads to the formation of sigmoidal fibers. In this example, the opening is first perpendicular to the vein wall, and then is oblique to the vein. Because of the locus of vein fiber precipitation, (c) antitaxial veins and (d) syntaxial veins have different shapes.

presence of a median line helps you in recognizing this important kinematic difference, but uncertainty about the interpretation can remain.

7.8 LINEAMENTS

A geologic **lineament** is a linear feature recognized on aerial photos, satellite imagery, or topographic maps. Lineaments generally are defined only at the regional scale; that is, they are not mesoscopic or microscopic features. Structural lineaments, meaning ones that are a consequence of the localization of known geologic features, are defined by structurally controlled alignments of topographic features like ridges, depressions, or escarpments (Figure 7.28). They may also be manifested by changes in vegetation, which is, in turn, structurally controlled. Most lineaments are the geomorphologic manifestation of joint arrays, faults, folds, dikes, or contacts, but some remain a mystery and do not appear to be associated with obvious structures.[6] You should maintain a healthy skepticism when reading articles about lineaments that do not confirm imagery interpretations with ground truth. Some "lineaments" that have been described in the literature turn out to be artifacts of sunlight interaction with the ground surface, and thus do not have geologic significance. However, study of true structural lineaments often provides insight into the distribution of regional structural features, ore deposits, and seismicity.

7.9 CLOSING REMARKS

In this introduction to the rapidly evolving science of joint analysis, we hope to have conveyed not only descriptive information about the structures, but also a sense of the controversy surrounding their interpretation. One of the common questions that students ask

[6]Except perhaps ancient landing strips for extraterrestrials.

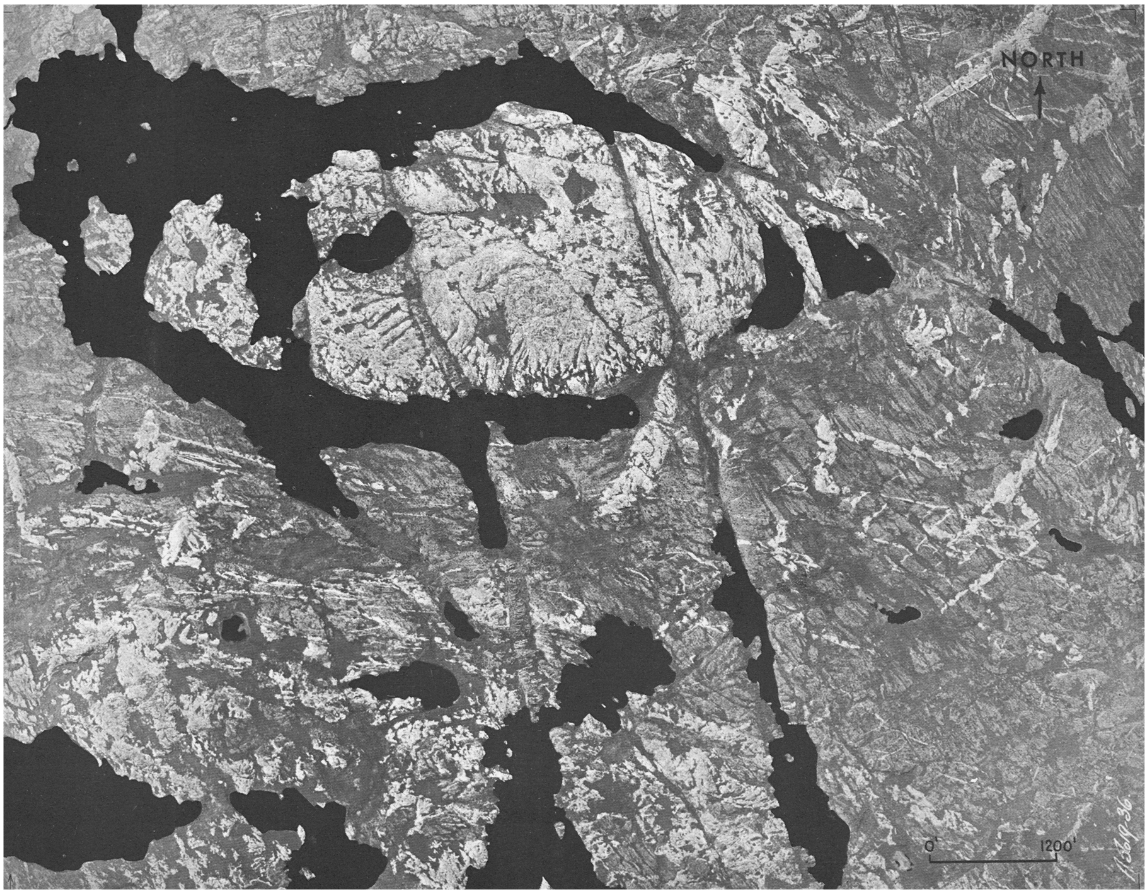


FIGURE 7.28 Aerial photograph of the Duncan Lake area (Northwest Territories, Canada), showing lineaments and structural control of topography. Scale is 1:14,500.

when beginning a field mapping project is, "Should I pay attention to the joints and veins?" An astute advisor might answer the question philosophically, with the words, "That depends."

If the purpose of the map is to define variations in permeability, or the location of faults, or the distribution of ore deposits, or the meaning of satellite-imagery lineaments, or the composition of fluids passing through the rock during deformation, then the joints and veins in the area should be studied. Perhaps you will find an interpretable variation in joint intensity within your map area, even if there is no systematic pattern to the jointing. Joint study is particularly important in studies of rock permeability, because the rate of fluid flow through joints may exceed the rate of flow through solid rock by orders of magnitude. Joints may make otherwise impermeable granite into a fluid reservoir, and may provide cross-formational permeability that permits oil to leak through an otherwise impermeable seal, or allow toxic waste in groundwater to leak across an aquitard. If the purpose of your mapping is to develop an understanding of paleostress in the region, then it may only be worthwhile to study joints if you can identify systematic sets. If the purpose of your study is to develop an understanding of stratigraphy in the map area, of the history of folding in a high-grade gneiss, then joint analyses probably won't help you very much and you shouldn't spend too much time looking at them. Of course, this advice might change in the coming decade, as sophistication in our understanding of the meaning of joints continues to grow. "That depends" remains a good answer, therefore.

Joints, by definition, are rock discontinuities (fractures) on which there has been no shear. In the next chapter and the final chapter on brittle deformation, we shift our focus to fractures on which there clearly has been shear; that is, faults. As you study Chapter 8, keep in mind the features of joints that we just described, so that you can compare them with the features and meaning of faults.

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